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CORRECTIONS AND EMENDATIONS

Contributors have been invited to send corrections and emendations to be made in their papers, and the issues have been scanned with some care. Corrections are as follows:

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Page 1773, equation 9 lines from bottom of page, extend square root symbol across entire equation

Page 1778, line 11, for last two sentences in paragraph, substitute the following: "The possibility that different amounts of heavies may come from different sources was discussed and illustrated by Krynine in the second of two papers (*Age of till on palouse soil from Washington*, Am. Jour. Sci., 5th ser., vol. 33, p. 205–216, 1937). For most investigations an assumption of equal weights will not be valid."

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Page 194, 3 lines from bottom, *for* Pl. 5, fig. 3, *read* Pl. 6, fig. 3.

Page 400, caption for Figure 15, *for* Co-coronoid, *read* Co-coracoid.

Page 511, line 8, *for* N. 35° E., *read* N. 35°. W.

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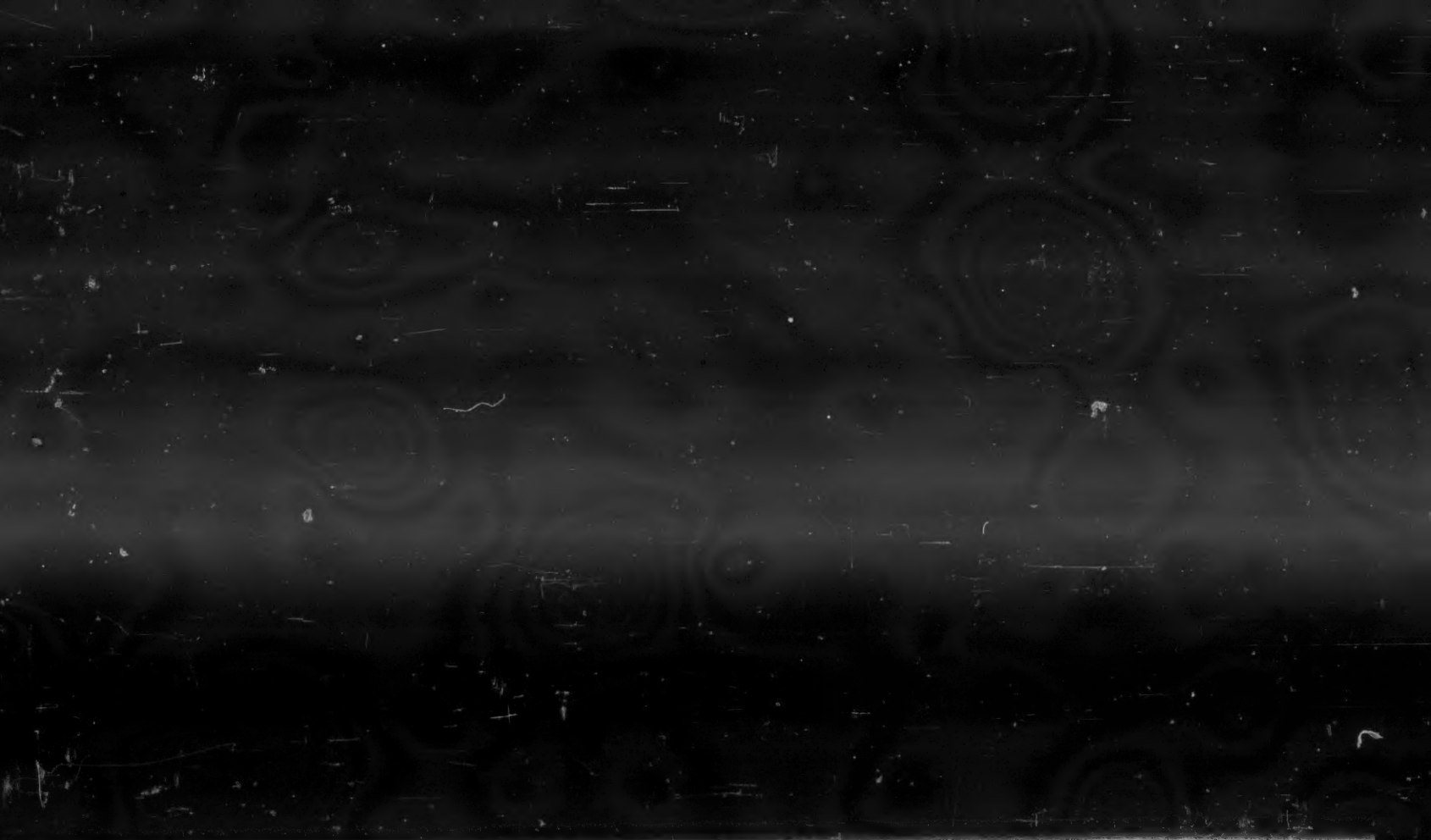

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Stratigraphy and Structure of West-Central Vermont. By Wallace M. Cady

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LOWER MISSISSIPPI VALLEY LOESS

BY RICHARD JOEL RUSSELL

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ABSTRACT

Lower Mississippi Valley loess resembles in all essential respects that of the Rhine Valley and other parts of Europe. Its field relationships preclude the possibility of eolian, lacustrine, fluvial, or other direct sedimentary origin. Typical loess grades upslope into parent material, from which it has been differentiated by a process here called loessification. Initial parent materials are terrace deposits physically similar to backswamp clays of the Recent Mississippi River. Parent materials weather into brown loam that creeps downslope, accumulating in greatest thickness in valleys and as mantles of bluffs. During loessification, carbonates accumulate, the size of particles becomes restricted mainly to 0.01–0.05 mm., snails are incorporated, and other loessial characteristics appear. More widespread development of loess east of the Mississippi results from wider areas of Pleistocene backswamp deposits. Loess occurs along major Mississippi tributaries, on such residual eminences as Crowleys Ridge, Sicily Island, and various hills west of the river in Arkansas and southeastern Missouri. Under some exposure conditions deloessification occurs, carbonates leach, coarse granules disintegrate, loessial characteristics disappear, and a type of brown loam is the end product. If a sharp distinction be made between loess and loesslike materials, it appears that the origin of loess in other regions is similar to that in the lower Mississippi Valley.

INTRODUCTION

When the writer began his lower Mississippi Valley field work, in 1928, he regarded the loess problem as settled and accepted the eolian hypothesis of origin. This belief was somewhat shaken when he found gravels and stratified beds in material he regarded as loess. During the next 10 years accumulating evidence favored fluvial origin, and investigations made during three trips to Europe tended to confirm that idea. At the annual meeting of the Association of American Geographers in 1936 he presented a short paper advocating fluvial origin, assigning the physical change from floodplain silt to typical loess to processes taking place on bluffs, where the deposits were being subjected to environmental conditions quite unlike those at sites of original deposition. Within more recent years evidence has accumulated that fluvial deposits do not change to loess *in situ*. They must first experience weathering and colluvial transportation.

Apparently a problem has arisen as a result of calling too many things loess. The necessity for a strict and wholly physical definition is inescapable before the question of origin may be considered seriously. After a definition of this kind is adopted it is possible to relegate many types of material to the general category, "loesslike deposits," and thereby remove many diverse, and often obvious, origins.

ACKNOWLEDGMENTS

To Professor H. N. Fisk the writer is indebted for many of the ideas advanced in this paper. Together we have examined in detail the deposits on Crowleys Ridge, many other localities west of the Mississippi, and outcrops along at least every main road and many of the byways in the loess belt of Kentucky, Tennessee, Mississippi, and Louisiana. To Professors W. Panzer, of Heidelberg, and Carl Troll, of Bonn, the writer is indebted for expert guidance to various loess localities in Germany, and to A. Penck for the term loessification.

A study of the literature has expanded gradually into a collection of abstracts and notes on more than 700 books and articles. The writer is indebted to Josephine G. Keller and Felix D. Richardson, of Louisiana State University, for their care in searching through complete sets of numerous American and foreign periodicals for titles of papers bearing on the problem, and to Marguerite M. Hanchey, Librarian in the School of Geology, for able assistance in the procurement of many items essential to the progress of the investigation.

For mechanical analyses the writer is indebted to Leo Hough, State of Louisiana, Division of Highways, and to Professor R. Dana Russell, Louisiana State University. Many helpful suggestions from Director H. V. Howe and Professor Chalmer J. Roy, School of Geology, Louisiana State University, have contributed to the progress of the study.

DEFINITION OF LOESS

The definition of loess should be rigid and descriptive,—devoid of hypothesis Materials resembling loess in one or more regards may be described as loesslike.

An examination of the literature reveals practical unanimity of opinion concerning

the meaning of *Löss* for several decades after its introduction as a scientific term in about 1821 (Scheidig, 1934, p. 58), or its English equivalent, introduced by Lyell (1834, p. 111). After the middle of the century an increasing number of loesslike materials were being included in the term. The present century has witnessed a strong tendency to introduce hypothesis of origin into the definition. Smith and Norton (1935), for example, define loess as

> "the sediment which is believed to have been produced by the grinding action of glaciers, deposited in the bottom lands from streams carrying water from the melting glaciers and then picked up and redeposited on the upland by the wind."

Todd (1918, p. 111) decried this tendency, but Tilton (1925) has insisted that eolian origin be made an essential part of the definition.

Observations appear irreconcilable to such a degree that over 20 different theories of origin have been proposed (Druif, 1927, p. 72–165). Interpretations are correct in many instances, but the diversity arises largely from the fact that many relate to material that is not loess at all.

No one seems to have formulated a precise, physical definition at an early date. A synthesis based on descriptions, however, indicates the intent to regard loess as homogeneous, unstratified, slightly indurated, porous, calcareous sedimentary rock consisting predominantly of particles of coarse silt size, ordinarily yellowish or buff in color, capable of splitting along vertical joints, and tending to stand in vertical faces. Everyone today seems to regard such material as "typical" loess, and it is doubtful that a better definition can be formulated. Quantitative limits might be added, but with caution. It is not safe to seek them in recent tabulations of mechanical or chemical analyses, such as appear in Scheidig (1934, p. 73–111), because they ordinarily include many samples of loesslike materials.

Most loesslike materials are excluded if a fair lime content be regarded as an essential characteristic. Scheidig regards as typical a calcium and magnesium carbonate content of 10–25%, though some lime-rich loess may contain 36%. Wailes (1854, p. 231) suggested 30% and over as typical in Mississippi. Keilhack (1920, p. 157) used 10–25% as typical. Calcium is the dominant carbonate and occurs as a cement throughout the entire mass. In addition it is present in concretionary nodules, root tubes, films concentrated along joints and faces, and in snail shells.

Sharp limitation in grain-size distribution is probably the most significant single diagnostic characteristic. Lyell (1834, p. 120) noted that it is far more homogeneous than alluvium. Leverett (1894, p. 111) observed that it contains far less fine material (< 0.005 mm.) than silt, and no coarse particles. Chamberlin and Salisbury (1885, p. 279) present detailed tabulations of grain sizes; their values are stated in millimeters but obviously refer to centimeters. Their counts show the predominance of particles in the fraction 0.01–0.05 mm., as do all subsequent studies. Keilhack (1920, p. 157) gives the following values as typical: 0.1–0.5 mm., 8–40%; 0.02–0.05 mm., 50–65%. In the writer's experience the single fraction 0.01–0.05 mm. ordinarily constitutes at least 50% (by weight) of a sample and in many cases amounts to 75%. Table 1 is representative of loess and some related materials.

TABLE 1.—*Grain-size distribution in samples of loess, leached loess, and a loesslike material*

(In per cent)

Sample (by weight) finer than	Loess			Leached loess	Loesslike
	1	2	3	4	5
Mm.					
0.5	99	98	100	100	100
0.4	99	98	100	100	100
0.3	99	98	100	100	99
0.2	98	98	100	100	96
0.1	98	97	99	100	84
0.09	97	97	98	100	83
0.08	96	96	98	100	82
0.07	95	95	98	100	81
0.06	93	94	97	99	79
0.05	90	91	96	96	77
0.04	82	86	90	90	75
0.03	64	76	78	81	71
0.02	33	50	59	54	60
0.01	15	19	30	23	38
0.009	14	16	28	20	34
0.008	12	14	25	18	32
0.007	11	11	22	15	29
0.006	10	9	19	13	25
0.005	9	8	16	11	21
0.004	8	6	14	9	18
0.003	7	5	11	8	16
0.002	6	3	8	5	11
0.001	4	1	6	2	8

Location of samples: 1. Haarlass, northside Neckar Valley, 1.5 km. above Alte Brücke, Heidelberg, Germany, 25 meters above river.

2. Five miles southeast of Natchez, Miss., U. S. 61, at junction between Woodville and Kingston roads, 4.5 miles east of Mississippi Valley bluffs, on divide between St. Catherine and Second creeks.

3. Center of Sicily Island, north-central Louisiana, west side of Mississippi Valley, on small stream leading to Ouachita River.

4. A few inches directly above sample 2.

5. Pleistocene terrace remnant 10 miles south-southwest of Delhi, La., 1.5 miles west of Lamar, west side of Mississippi Valley, northeastern Louisiana.

If the percentage of material finer than 0.01 mm. be subtracted from that finer than 0.05 mm., the remainders in samples 1, 2, and 3 are 75%, 72%, and 66%. This remarkable degree of sorting is also exhibited by leached loess sample 4, with a remainder of 73%. Although the outcrop represented by sample 5 exhibited so many loessial characteristics that many persons would identify it as loess, the degree of sorting is less perfect in several regards, and only 39% falls within the size limits appearing critical in the definition.

The definition should include the following essential characteristics: Loess is unstratified, homogeneous, porous, calcareous silt; it is characteristic that it is yellowish or buff, tends to split along vertical joints, maintains steep faces, and ordinarily

contains concretions, and snail shells. From the quantitative standpoint at least 50%, by weight, must fall within the grain size fraction 0.01–0.05 mm., and it must effervesce freely with dilute hydrochloric acid.

In the discussion to follow the word loess is used only to describe materials that meet all parts of the definition. Leached loess is used only to describe material immediately overlying unaltered deposits, differing only in absence of calcareous content, which unquestionably was at one time loess. Other similar materials are in many cases described as loesslike.

LOESSLIKE SEDIMENTARY ROCKS

Many sedimentary rocks closely resembling loess in one or more regards should be called loesslike. In literature they are ordinarily designated by such terms as river-loess, lake-loess, marine-loess, loess-loam, noncalcareous loess, or stratified loess. The German language is particularly rich in such expressions. With regard to position it offers: *Deckenlöss, Höhenlöss, Plateaulöss, Flankenlöss, Beckenlöss, Hanglöss, Gehängelöss, Flusswinkellöss, Bogenlöss, Terrassenlöss, Steppenlöss, Luvlöss, Leelöss*, and others, and with regard to origin: *Primärlöss, Sekundärlöss, Dejektiver- or Schwemmlöss*, and *Inundationslöss*. Age distinctions are recognized, such as: *jünger Löss, älterer Löss, fossiler Löss*, and *rezenter Löss*. There are also such terms as *Lösslehm, entkalker Löss*, and *vertoner Löss*. French is also rich in loess terms, and many other languages exhibit the same tendency, as may be appreciated by the fact that almost every expression listed has its direct equivalent in English. Some of these words refer to loess, but most have originated as the result of uncritical definitions, extended beyond all reasonable limits.

To counterbalance the tendency toward inclusion of altogether too many loesslike materials in the definition of loess many writers use such terms as typical, true, genuine, pure, and, in all, over 50 similar adjectives in the more common European languages. This confusion disappears when it is realized that so many references to loess actually mean loesslike materials.

The residual weathering of many types of silt and clay yields loesslike products. Jongmans and Van Rummelen (1938, p. 120) report Cretaceous bryozoans and echinoids in loess in southeastern Netherlands. The writer has visited the field locality and confirms all conclusions but one—the material is actually loesslike. Quaternary alluvial deposits are especially prone to assume loessial characteristics during weathering on outcrops providing good drainage. The writer recently regarded such materials as immature loess (1938, p. 75), failing to realize that only under special conditions, where colluvial transportation occurs, can any of them become loess.

The physical characteristics most commonly responsible for confusion are color and tendency to fracture vertically. Obscure stratification, superficial textural similarity, and impalpability also lead to confusion.

Terrace materials, to a greater degree than other types of deposits, assume many loessial characteristics. The writer estimates that over half of the American literature on loess actually refers to loesslike terrace silts. This confusion exists elsewhere. Many of the photographs by Willis (1907) and the excellent sketches and photographs by Barbour (1929) clearly show flat-surfaced alluvial terraces in China.

The areal extent of loess deposits has been grossly exaggerated by widespread

inclusion of loesslike materials. Tillo (1893) considered 4% of the land surface of the earth covered by loess. For the entire latitudinal zone between 20°S. and 40°S. his estimate was 13%, for South America, 10%, and for North America, 5%. Keilhack (1920) raised the estimate to 9.3% for the earth's land area. Detailed distribution maps (Scheidig, 1934, p. 7–25; Jenny, 1941, p. 58) include nearly all areas where alluvial terraces are extensively developed. In the lower Mississippi Valley the areal extent of loess is an extremely small fraction of the belt ordinarily so designated. The writer has found the same to be true in the Rhine, Danube, Rhone, and other valleys in Europe.

Lower Mississippi Valley literature is difficult to interpret because loesslike deposits may be called loess by one writer and by some other name, such as brown or yellow loam, by another. In addition brown loam has been used for dozens of distinctly different materials. The weathering of many Tertiary formations yields a number of residual products that are ordinarily called brown loam. These may be individually distinct, and all are quite different from brown loam that has weathered from Pleistocene terrace deposits. The weathering of loess yields still another type of brown loam. The term is also used for various soils. As a rule all such materials are either noncalcareous or only faintly so. They fail to show diagnostic sorting characteristics of loess. Some are decidedly loesslike at the outcrop. Others are so different that they have never caused confusion.

The least justified terminological practice is that of calling a material loess because of known or suspected eolian origin. Deposits of accumulating dust ordinarily bear no physical resemblance to loess, and few deserve to be called loesslike. Davis (1905, p. 62) and Huntington (1907, p. 103) applied the name to dunelike accumulations of powdery white dust in Turkestan. R. W. Pumpelly (1908, p. 271) experienced dust storms that led him to the conclusion that he was in a region of "still living loess." Hobbs (1931, p. 383) speaks of the difficulties in walking in deep loess freshly deposited on tundras. Such statements are quite in contrast to those of Schokalsky (1932, p. 85), who clearly distinguishes between loess and the dust covering it in the Salt Range of India. Berg (1932, p. 133) has observed that much of the dust so widespread in Asia is "the result of man's activities, being a product of deflation of the upper horizons of soils developed upon, and from, loess . . . but in no way can anything like loess be obtained as a result of its subsequent deposition." Clapp (1920) and Cressey (1934, p. 75) clearly relate dust in the air of northern China to the cultivation of fields or to the presence of extensive loess deposits.

Redeposited loess is a confusing term. In the lower Mississippi Valley it ordinarily means silty terrace deposit and implies evident stratification. This expression should be restricted to materials of known genesis and used only in a strict, wholly literal sense. About the only examples are such things as miniature fans associated with gully dissection.

THEORIES ON THE ORIGIN OF LOESS

Theories on origin run almost the entire gamut of geological possibilities. Direct volcanic deposition was postulated by Howorth (1882) and others. I. C. Russell (1897, p. 59–63; 1901, p. 81–83) regarded loess as a residual deposit from weathered

basalt in eastern Washingron. An examination of his localities reveals the presence of loesslike material that is indeed residual, with remnants of veins extending through it, toward the surface. Keilhack (1920, p. 161) concluded that loess is of cosmic origin, being derived from a dust ring resembling those of Saturn. A strange cause of Pleistocene glaciation is involved in his hypothesis. At the other extreme is the conclusion of Campbell (1899) that loess was derived from animal and vegetable matter. Orton and others proposed the idea that earthworms and burrowing animals created it by bringing fine materials to the surface (G. F. Wright, 1890, p. 104; Savage, 1915, p. 106). Druif (1927) and Scheidig (1934) give more complete statements of various theories, but they overlook recent Russian literature.

Ehrlich (1848) originated the eolian hypothesis according to some writers, and it may be that his study of a dust storm that covered much of central Europe, an area in which loess is relatively abundant, was the basis for this belief, but an examination of several of his writings revealed no direct statement of the hypothesis.

Virlet d'Aoust (1857) deserves credit for the eolian hypothesis, having definitely prescribed it for deposits in Mexico. Freudenberg (1909, p. 274) visited the localities in question and confirmed the identification of "typical yellow loess."

Credit for the eolian hypothesis is ordinarily given to Richthofen, who popularized it and is responsible for its vogue today. The original reference is a report on the Provinces of Honan and Shensi, written in Shanghai in 1870.

American geologists did not receive the eolian hypothesis well. Todd (1878), Child (1881), Winchell (1884, p. 263), Chamberlin and Salisbury (1885, p. 287), and Dana (1895) attacked it vigorously. Fluvial or lacustrine origin were the theories then in vogue. Among notable adherents to the former were Lyell (1834; 1838), Binney (1846), Dumont (1852), D. D. Owen (1852), Newberry (1889), Upham (1892), Todd (1897, p. 50), and, more recently, Fuller and Clapp (1903), L. A. Owen (1905), and G. F. Wright (1921), Lacustrine origin was favored by R. Pumpelly (1866), Hayden (1872), Call (1882), Mcfarlane (1884), Warren (1878), and Witter (1892b). Chamberlin and Salisbury (1885) thought both origins probable. Salisbury (1892, p. 317) restricted deposition to waters of glacial origin. Near coasts estuarine origin was favored by Gordon (1892), Hopkins (1872, p. 137), and Hilgard (1874). Lyell favored fluvial origin on the whole, but where valleys widened he pictured deposition in bodies of relatively still water. Thus, in the upper Rhine Valley loess was deposited in a "pre-existing basin or strath, bounded by lofty mountains" (1863, p. 330). Marine origin, advocated by Kingsmill (1871) and Skertchly and Kingsmill (1895), attracted few adherents. Some kind of aqueous origin explained loess according to most geologists for nearly 3 decades after Richthofen's eolian origin was advanced.

In the Central States there had long been a feeling that the lacustrine theory was unsatisfactory. Deposits in Nebraska were some 2000 feet higher than those along the Mississippi. Satisfactory boundaries of lakes could not be demonstrated. Shimek's (1897) patient research emphasized the terrestrial character of the fauna. Localities were too closely spaced and too numerous to permit lacustrine origin (Shimek, 1903; Chamberlin and Salisbury, 1907, p. 410; Leverett, 1932).

R. Pumpelly was so impressed by Richthofen's thesis that he retracted a whole series of observations and conclusions, advanced in support of lacustrine origin in

1866 as a result of his own explorations in China (1879; and many restatements, *e.g.*, 1908; 1918, p. 611ff.).

European sentiment also swung toward the eolian hypothesis. Land mammals were being found in material regarded as loess, and many exhibited dry-climate affinities, suggesting that deposition of dust from Asiatic or African deserts had occurred during inter-glacial stages.

It was easier for Americans to abandon the lacustrine than to accept the eolian hypothesis. Some favored fluvial origin, but their evidence, on the whole, suggests stratified, loesslike deposits. Barbour (1929, p. 70) is correct in stating that such materials are "one cause for the slow death of the aqueous theory."

Chamberlin and Salisbury (1885, p. 287) proposed a combination theory invoking both fluvial and eolian agencies, but after many pages of discussion rejected it in favor of aqueous origin. Chamberlin (1897) later regarded the combination theory more favorably, and it was eloquently restated by Keyes (1898). Sardeson (1899) advocated eolian origin. Reports from China and Turkestan (Obrutchev, 1895; Huntington, 1907) to the effect that loess was being formed by deposition of dust appeared to clinch all arguments in favor of eolian origin.

Today most Americans appear to favor the idea that rivers transported fine glacial debris to broad flood plains, from whence it was picked up by winds and deposited on, or near, adjacent bluffs, particularly along the eastern sides of valleys. This is essentially the combination theory of Chamberlin and Salisbury (1885) and has been well stated many times (Fairchild, 1898; Willis, 1907, p. 249; W. B. Wright, 1914; Emerson, 1918; Fenneman, 1938, p. 97; Lobeck, 1939, p. 377). Todd (1897) and later L. A. Owen (1904; 1926) attacked it vigorously. Recently Flint (1941, p. 27) stated that 3 decades have elapsed since "the last serious opposition to the eolian view was voiced."

Minor theories deserve only brief notice here. Some run along such obsolescent lines as a marine catastrophe involving waters at least 700 feet deep (Hibbert, 1834). Others invoke snow drifts (Davison, 1894), alternate freeze and thaw (Kerr, 1881; Wood, 1882), or residual weathering (Peters, 1859; Linstow, 1910; Rummelen, 1923).

The close relationship between loess and Quaternary terrace deposits was noticed by Fallou (1867), who differentiated it from brown loam on the basis of carcareous content. Harris and Veatch (1899, p. 177) regarded it as "a local development of the loam." Winchell (1903, p. 141) regarded it as a direct and contemporary variation of till. Oefelein (1934) noted that weathering of "gray loess" produces a buff color. Observations along these lines have been receiving increasing attention in Russia, where they have been summarized by Berg (1932) in a paper that should not be overlooked.

IDENTIFICATION OF LOESS IN THE LOWER MISSISSIPPI VALLEY

The word *Löss* found its way from popular into scientific terminology in the Rhine Valley during the early part of last century. Detailed faunal studies by Braun (1847), started in the vicinity of Frankfurt-am-Main, established the presence of identical materials in various parts of western and central Europe.

With a background of familiarity with the Rhine Valley deposits, Lyell visited Natchez, Mississippi, in 1846 and later (1847), with reference to the river bluff section, wrote:

"The yellow loam at the top bears a singularly close resemblance to the fluviatile silt, or 'loess,' as it is termed, of the valley of the Rhine, between Cologne and Basle, and, like it, contains abundance of freshwater and land shells . . . the loam, unsolidified as it is, retains its verticality, as is the case with its counterpart, the loess of the Rhine."

After visiting many of the classic Rhine Valley localities, as well as deposits elsewhere in Europe, the writer agrees that loess is a distinctive material occurring at many places. The deposits at Natchez and other places in the lower Mississippi Valley differ in no significant manner from those in the Rhine Valley. Physical similarity has been demonstrated in Table 1. Faunal similarity will be discussed later.

FIELD RELATIONSHIPS OF LOWER MISSISSIPPI VALLEY LOESS

DISTRIBUTION

The conventional idea of loess distribution south of the Ohio River pictures a more or less general cover about 15 miles wide, extending eastward from bluffs facing the Mississippi alluvial valley. Geological maps show this belt (Stephenson, 1928, Pl. 2). Physiographers designate it as Loess, or Bluff, Hills (Fenneman, 1938, p. 80–81). Pedologists outline the same region by showing generalized distribution of soils of the Memphis series (Marbut, 1935). These soils are also present on Crowleys Ridge and various hills west of the river, where loess also occurs.

The fact has not been stressed sufficiently that only a minute portion of the Loess Hills belt is actually covered. Outcrops are limited to slopes. Exposures appear thick on bluffs, but deposits are thin on most slopes and are wanting on upland flats.

A detailed distributional map for the lower Mississippi Valley has never been made. Its preparation would be extremely tedious, involving many decades of intensive field work. On a scale such as 1:500,000 few outcrops would appear other than as narrow lines. Many would be dots or small dashes. Lines would be most continuous along bluffs facing the Mississippi flood plain and along some of the larger tributaries to the east. Dashes would outline the flanks of ridges and spurs in deeply dissected territory and would be most numerous in southern Mississippi. In northern Mississippi few would be found more than half a mile from the flood plain. Crowleys Ridge, particularly toward the south, would be one of the most conspicuous areas on the map. With the possible exception of a small region at Vicksburg it is unlikely that any continuous outcrop has a width of as much as 1 mile. Although outcrops are rather continuous along the bases of major bluffs, loess commonly fails to reach the summits of the slopes above them.

The belief that loess is heaped up to greatest elevations on bluffs and that it thins eastward, with decreasing elevation, is unsound. Slopes are generally toward the Mississippi Valley. Shimek (1902) claimed eastward-sloping loess surfaces at Natchez, but the topography is related to dissection by tributaries of St. Catherine Creek, a stream that winds around the town in a deep valley. Loess along that

stream occurs at elevations both higher and lower than the tops of the bluffs along the Mississippi, as it does along other main streams in that vicinity. It seems to be agreed that depositional slopes lead away from streams north of the Ohio River (Fenneman, 1938, p. 508), but the absence of corresponding features to the south suggests that it might be well to re-examine the northern deposits from the viewpoint of detecting erosional origin of the topography.

Deposits are related to slopes, not compass orientation. They are equally thick on either side of ridges exhibiting topographic symmetry.

Though the most spectacular development occurs along bluffs east of the Mississippi River flood plain it is by no means restricted to that location. Clendenin (1896) found it on both sides of the valley. Todd (1897, p. 47) correctly described it as "thickest near streams on both sides alike." G. F. Wright (1904) expressed similar views, noting, however, that in the upper Mississippi Valley, especially in the Driftless Area, the upper limit is ordinarily higher along western sides of valleys (1911, p. 413). On Sicily Island, west of the valley, in Louisiana, it was identified by Clendenin (1897), Harris (1902), and Veatch (1905). A mechanical analysis of this material is given in Table 1. Loess on Crowleys Ridge was studied in detail by Salisbury, Call (1889), and Shimek (1916). Hershey (1895) noted it in valleys leading from the Ozark Plateau. The larger streams across Mississippi are flanked by loess at places. Wailes (1854, p. 213) found it along the Homochitto, Big Black, and Yazoo rivers. Three years later Harper (1857, p. 254) repeated the observation.

The writer's field work demonstrates that the distribution of lower Mississippi Valley loess depends upon two main factors: (1) slopes, and (2) specific types of Pleistocene terrace deposits, from which it can be derived.

STRATIGRAPHIC RELATIONSHIPS

From the stratigraphic standpoint lower Mississippi Valley loess occurs only as mantles leading upslope to outcrops of finer sediments of Quaternary terraces. No deposit reaches an elevation equal to that of the surface of the highest terrace in its vicinity. On slopes it covers both Tertiary and Quaternary deposits and incorporates materials from these underlying beds.

It is not a geological formation in the technical sense of the term for it has no fixed stratigraphic position. Traced upslope it grades laterally into the upper part of any one of three different Pleistocene formations. It is not a soil for soil profiles in process of differentiation may be seen on its surface. It is parent material for soils, just as is alluvium or any other kind of rock. In many outcrops, however, it is separated by many feet from soil by deeply weathered material that may be described by the rather unsatisfactory term brown loam.

Where exposed on bluffs deposits may appear extremely thick. Such appearance is deceptive, being the result of mass movements that spread relatively thin sheets over slopes having considerable height. At Natchez loess is ordinarily regarded as having a thickness about equal to the height of the bluff, or about 200 feet, and on the road leading to the old ferry landing at Natchez-Under-The-Hill the outcrop is continuous almost to the base of the bluff. However, less than 1 mile to the south, in recent excavations immediately south of the Mississippi River Bridge, the bluff is

composed almost wholly of stratified terrace deposits, including cross-bedded gravels, sands, and silts, the loess cover being nowhere over 50 feet thick. Between these excavations and the river a thin mantle of loess has crept down the bluff. Equally impressive proof of creeping exists just north of Vicksburg and elsewhere.

The mantling of bluffs is not the only type of mass movement involved in giving an appearance of extreme vertical thickness. Bluffs everywhere are subject to conspicuous slumping, and the repetition of sections on slumped blocks is obscured by healing of fractures in an amazingly short time. Notable examples occur near Fort Adams, near the southern boundary of Mississippi, on the bluff not far south of the bridge at Natchez, and just north of Helena, near the southern end of Crowleys Ridge.

Literature often confuses actual thickness with vertical range. Richthofen's (1877) descriptions are carelessly worded in this regard. Though it appears doubtful that he actually wished to imply a stratigraphic thickness of 2000–2500 feet for sections along the Yangtze Valley, he is given credit for such values by many writers (R. Pumpelly, 1879, p. 133; Hay, 1914, p. 39; Moyer, 1932) and in many textbooks (following Le Conte, 1897, p. 584). Schmitthenner (1919, p. 313) gives the actual thickness as closer to 60–80 meters than to 600–800. L. A. Owen (1926) observes that the thickness is rarely over 150 feet, rather than 1500, stated by Richthofen. A misunderstanding that caused Upham (1891) to state that materials similar to loess have a thickness of 1500–2000 feet in parts of the Great Basin and San Joaquin Valley of California is traced to I. C. Russell's curious idea that adobe is the same as, or similar to, loess. This idea appears in Tarr and Martin (1914, p. 72) and is discussed by Lobeck (1939, p. 397). In the writer's experience loess is ordinarily less than 50 feet thick in a direction normal to its surface. Cores taken in vertical holes often exaggerate its true thickness.

Many of the best exposures occur in faces of artificial cuts across ridges or spurs. The typical appearance is pseudoanticlinal. Such exposures reveal the close relationship between distribution and topography. In many long cuts loess may be found only toward either end. In shorter cuts it ordinarily thickens down the slopes determining the limbs of the pseudoanticline, a typical case being shown in Figure 1 of Plate 1. At the crest of the spur a thin layer of soil is being differentiated from leached loess that thickens toward the left and may be identified by light color. A similar thickening occurs in the darker loess below. In the central part of the cut stratified gravels form the core of the spur. These are abruptly truncated toward either limb of the pseudoanticline. In the lower parts of the exposed limbs there is a notable amount of gravel admixed in the loess, the thickness of the mixed material increasing downslope. Gravels in the core of the spur are stained deep red, and individual stones lie with flat sides down, in orderly stratigraphic arrangement. Gravels admixed in the loess are gray; stones have random orientation and exhibit no evidence of stratigraphic continuity. Gradation between the gray gravel zone and loess is sharp toward the crest and becomes increasingly hazy down the limbs. Snail shells are abundant and entire throughout the loess, even where much gray gravel is admixed. They occur neither in the red gravels nor in the leached loess.

In practically all cuts where pseudoanticlinal relationships are exhibited the pitch of the axis is relatively steep, and it likely lies about normal to the bearing of the road.

The exposure on one side of a cut is quite different from that on the other, so that one may observe, in a three-dimensional manner, that loess thickens down the slope of the spur. The thickening at right angles to the road is in many cases more significant than that appearing on the limbs of the psuedoanticlines.

RELATION TO PLEISTOCENE BACKSWAMP DEPOSITS

There is a distinct correlation between occurrences of loess in the lower Mississippi Valley and the presence of backswamp deposits in the associated Pleistocene terrace formations. The terrace formation responsible for most of the loess is the Montgomery, but notable amounts have been derived from the older Bentley, and some from the oldest Pleistocene Williana formation.[1]

Lower Mississippi Valley Pleistocene terraces reveal topographic details exactly like those of the active flood plain. Fisk (1940, p. 68–86) identified natural levees, meander loops, point streams, and other meander-belt characteristics on the Prairie terrace. More recent studies, based both on aerial photographs and field work, show similar details on the older surfaces. The conclusion is inescapable that each of the major terraces developed exactly as has the Recent flood plain, as the result of an alluvial filling of a pre-existing valley system to a level determined by the Gulf of Mexico.

The relative flatness of the flood plain today is related to deposition alone. Lateral erosion has played no part whatever in its development. A gradual shifting of the Mississippi River from one side of its valley to the other is a physical impossibility. The channel is restricted to a meander belt, within which it shifts but from which it rarely escapes. A meander loop may be accentuated only to a definite limit before it is cut off, or chuting occurs, with the result that the channel again finds itself centrally located in the meander belt. Deposition of coarsest load is restricted to meander belts, where complex systems of natural levees are formed and elevations are raised notably above those of backswamps between such belts. Contrasting elevations in various parts of a flood plain ultimately lead to meander-belt instability and diversion of the main channel to an entirely new course downstream. Lower territory is present in distant backswamps, so that a new course is displaced widely from the old. The last major shift of the Mississippi River in Louisiana resulted from diversion near the Arkansas line. A meander belt flanking the extreme western side of the valley was abandoned in favor of a course along the extreme eastern side. The eastern meander belt has now been alluviated to an extent where the river is unstable, and the most likely course for the future lies along the low backswamp region in mid-valley.

The width of the present Mississippi River meander belt south of Baton Rouge, Louisiana, is about 10 miles. That of the Teche-Mississippi, or last western course, in the same latitude is also 10 miles. Between them is the Atchafalaya Basin back-

[1] Louisiana terminology will be used in this discussion. Fisk (1938) recognized four terraces and named them Prairie (lowest and youngest), Montgomery, Bentley, and Williana. Fisk and the writer have traced them along both sides of the alluvial valley into Missouri and Kentucky. The terraces are of alluvial origin. Each is the surface of a deposit formed during an interglacial stage as fill in a valley system cut during the preceding glacial stage (Russell, 1940). Fisk (1940, p. 175-183) later used the same four names for the formations beneath the respective surfaces. They are the equivalents of the European Monastrienne, Tyrrhénienne, Milazzienne, and Sicilienne, as developed along the southern Rhone Valley.

swamp, from 28 to 36 miles wide. The typical surficial deposit in this backswamp is calcareous clay 50 feet or more in depth. Floodplain clay deposition is not restricted to the backswamp. The typical deposit within loops of a meander belt is also clay, and heavy clays fill parts of many abandoned channels. Silt and coarser sediments are restricted mainly to natural levees and therefore to strips of from 1 to 3 miles wide on either side of channels. The sedimentary complexity along meander belts contrasts with simplicity in backswamps. An index of the complexity is the density of the channel scar pattern within a meander belt. The relatively stable channels toward the coast display few channel scars in their narrow meander belts. North of Baton Rouge the pattern is complex and an intricate system of sinuous, thin bands of coarser deposits crisscross similar bands and irregularly shaped masses of clay originally deposited along natural levee flanks and within their loops.

South of Natchez, in the latitude of the mouth of Homochitto River, the present Mississippi River meander belt is 10 miles wide. The Catahoula backswamp to the west is about 30 miles wide. North of Vicksburg, in the latitude of Lake Providence, Louisiana, the meander belt is 11 miles wide. The lower Yazoo delta backswamp is here about 35 miles wide, and an additional width of 10 miles of backswamp exists west of the river. In these examples the meander belt occupies but one fourth or one fifth the width of the flood plain, a proportion fairly typical where the flood plain is wide. In constricted areas the meander belt may occupy the entire valley width, as at Cape Girardeau, Missouri, or, nearly so, as at Helena, Arkansas, east of Crowleys Ridge. In the Latitude of Memphis, Tennessee, the meander belt is about 10 to 12 miles wide, and the backswamp to the west is about 40 miles wide.

This discussion of Recent flood plain is an essential part of the explanation of lower Mississippi Valley loess distribution because floodplain features are duplicated on all Pleistocene terrace surfaces, and their details correlate with the presence or absence of loess deposits today. Two points deserve emphasis: (1) backswamp deposits characterize wider parts of the flood plain, and (2) limited areas of similar materials occur within meander belts.

The most clayey parts of flood plains are the backswamps because their deposits are formed from only the finest mechanical sediments, brought in at times of flood, and they are the most organic. They are the most calcareous because they accumulate in basins where calcium carbonate and other soluble substances are concentrated. Typical backswamp waters are deeply stained by iron carbonates. Spoil from borings, ditches, or dredgings in backswamp deposits is invariably rich in concretionary materials. These nodular, lime-rich growths are not characteristic of silts along meander belts.

The most conspicuous deposits of loess in the lower Mississippi Valley grade upslope and are transitional into Pleistocene backswamp deposits. Where Pleistocene meander belts are well developed on the terraces loess is absent or poorly developed. The amount of loess present in a given locality may be predicted with amazing accuracy from an examination of terrace drainage patterns. The Montgomery terrace in northern Mississippi, south of Memphis, displays typical Pleistocene meander-belt patterns. Many loops are now entrenched or indicated only by paired streams and similar features. An examination of the Crenshaw, Mississippi, War Department

quadrangle will reveal a typical example. Here there is little loess. In central and southern Mississippi, where the Montgomery terrace displays widespread and typical backswamp drainage patterns, loess is abundant.

South of the Ohio River, loess appears to be confined to Pleistocene equivalents of the Mississippi Valley. Somewhat similar material has been reported near Natchitoches, Louisiana, but the writer has been unable to find typical deposits there or elsewhere along Red River. It is absent along Pearl River and other streams outside of Mississippi River drainage. This relationship may reflect the mineralogical or petrological nature of materials carried by rivers. The writer is more inclined to believe that it is related to widths of valleys, the Mississippi only having a flood plain wide enough to permit the accumulation of sufficient volume of backswamp deposits.

LOCALITIES SHOWING GRADATIONAL RELATIONSHIPS

Gradational relationships between loess and Pleistocene backswamp deposits may be noted almost wherever loess occurs in the lower Mississippi Valley. The following examples are cited not so much on the basis of individual excellence as on geographical distribution and accessibility. Following each locality heading is the name of the U. S. War Department, Corps of Engineers, quadrangle covering the area. These maps are obtainable from the Mississippi River Commission, Vicksburg, Mississippi. Highways are identified by numbers.

Hickman, Kentucky (Hickman, Missouri-Kentucky-Tennessee): Excellent exposures occur south of town in cuts along State Road 309, east of U. S. 94, leading toward Brownsville, both north and south of the railroad crossing. The material is calcareous, contains many snail shells, and is overgrown by moss. The color varies from ash to light buff. Nodular concretions are abundant. Lignitic Tertiary beds underlie the loess in the exposure next north of the railroad. Traced southward, upgrade for 0.5 mile, there is a complete transition from loess to terrace clays and silts that are noncalcareous, contain no shells, and lack moss cover. A similar transition occurs northward from State Road 309 into town, but the presence of houses and artificial disturbances render it more difficult to follow in detail. Along the bluff just north of Hickman terrace deposits cover Tertiary beds at elevations well toward the top of the bluff, and loess in this vicinity is only a thin mantle.

Near Tiptonville, Tennessee (Reelfoot Lake, Tennessee-Missouri-Kentucky): Excellent outcrops of loess occur along State Road 21, in Glady Hollow, 6 miles east of Tiptonville, on the road to Hornbeak. The material is calcareous, contains many snail shells, concretions reach large sizes gravels are abundant at lower elevations, and a mossy cover is present on outcrops. Tertiary beds and terrace deposits, including bedded gravels, underlie the loess. After leaving the main flood plain the road enters a valley where exposures may be seen for about 1.2 miles. A grade about 0.2 mile long then brings the road to a widespread terrace. The transition from loess to terrace deposit occurs on this grade, the upper materials being noncalcareous, lacking snail shells, concretions, and mossy cover. Similar transitions may be seen on either side of North Branch, on a secondary road leading from Hornbeak westward toward Wynnburg, and also where this road descends a steep grade to the flood plain.

Obion River, 6.5 Miles North-northwest of Dyersburg, Tennessee (Dyersburg, Tennessee): Excellent loess exposures occur on the grade just south of the river, on State Road 78, between Bogota and Nauvoo. This locality is particularly instructive. Flecks of lignite calcareous enough to effervesce with dilute acid are incorporated in the loess. In lower parts of the exposure large blocks of backswamp clay contain lignite. Mass movements have been so rapid that many of these blocks are fresh and sharply defined, though surrounded by loess. An exposure in a ditch along the bluff side of the road shows a gravel train, at places 4 feet thick, locally interbedded in snail-bearing loess. The transition into noncalcareous terrace materials devoid of snail shells occurs toward the top of the grade.

FIGURE 1. PSEUDOANTICLINE
Loess overlying gravel and underlying leached loess, 1.6 miles east on U. S. 84 from junction with 61 at Washington, near Natchez, Mississippi.

FIGURE 2. PARENT MATERIAL
Backswamp sediment exposed along road leading down from terrace surface about 1 mile north of Girls Camp Area, Shelby Forest, near Memphis, Tennessee.

FIGURE 3. COLUMNAR LOESS
Typical loess on bluff facing Mississippi alluvial valley on Herring Hill Road, about 4 miles northeast of Girls Camp Area, Shelby Forest, near Memphis, Tennessee.

TYPICAL EXPOSURES

FIGURE 1. PARENT MATERIAL
Alluvial terrace deposit exposed at summit of short grade leading down the grade to the flood plain west of Eudora, Mississippi.

FIGURE 2. TRANSITION TO LOESS
Colluvial material about 0.1 mile down the grade mentioned above. Material calcareous enough to effervesce with dilute acid is found only on joints, surface crusts, and along roots.

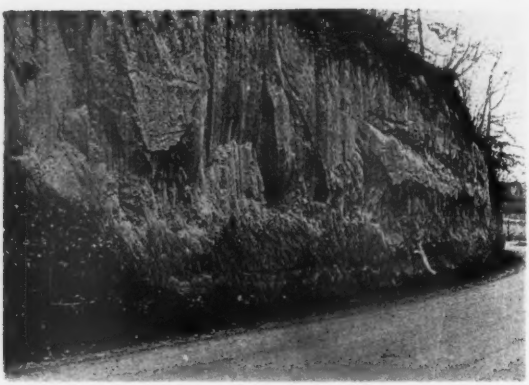

FIGURE 3. TYPICAL LOESS
Snail-bearing loess about 0.1 mile below the outcrop shown in Figure 2 of this plate, with Mississippi Valley flood plain in the distance.

TRANSITION TO LOESS

SHELBY FOREST, TENNESSEE (Jericho, Arkansas-Tennessee; Millington, Tennessee): Notable loess exposures occur about 15 miles north of Memphis along the bluff facing the flood plain and north of Shelby Forest. About a mile north of the Girls Camp Area a secondary road leads down a steep grade to a high bridge. The upland surface is a flat terrace. Where the road starts down the grade cuts on either side (Pl. 1, fig. 2) reveal typical backswamp deposits. About half way down the grade material that effervesces with dilute acid is restricted to coatings on roots and thin surface crusts in the vicinity of faint vertical joints. The transition to loess is gradual. Near the foot of the grade limy tubular coatings occur on roots, and calcareous crusts cover exposed surfaces. Typical, snail-bearing, calcareous loess, with many concretionary nodules, outcrops in the ravine below the road, immediately south of the bridge.

A similar section occurs about 3 miles northeast, on Herring Hill Road, north of the store by that name, on a steep grade leading to the flood plain. This road roughly parallels the bluff and hence displays considerable contrasts between exposures on its opposite sides. Loess is thicker and more calcareous on the downslope side of the cut and forms a relatively thin mantle overlying Tertiary and Quaternary deposits. Excellently developed columnar jointing occurs not only in the loess (Pl. 1, fig. 3) but in underlying Tertiary beds as well, as may be observed near the foot of the grade on the bluff side of the road.

BETWEEN BANKS AND EUDORA, MISSISSIPPI (Horn Lake, Mississippi-Tennessee-Arkansas): A narrow band of loess on bluffs, typical of northern Mississippi, occurs along State Road 3, west of Hernando, between Banks and Eudora. Similar sections may be seen along many minor roads leading from flood plain to terrace surface southward for some 30 miles (Crenshaw, Mississippi; Sledge, Mississippi). The terrace in this region is characterized by meander-belt drainage patterns and materials.

East of Good Hope Church, on State Road 3, the transition between loess and terrace deposits occurs in less than 0.3 mile, and no other loess is exposed eastward along the road. At the summit of the grade rising from the flood plain is typical terrace deposit, a noncalcareous clay that tends to weather in a pinnacly manner (Pl. 2, fig. 1). About half way down, some loessial characteristics appear, cuts stand vertically, and buff color appears (Pl. 2, fig. 2). Here the material effervesces along weakly developed joints and on root coatings. Toward the foot of the grade typical loess occurs (Pl. 2, fig. 3). All gradations between these materials are without sharp boundaries.

EAST OF GREENWOOD, MISSISSIPPI (Greenwood, Mississippi; Seven Pines, Mississippi): Sixteen loess exposures occur between the edge of the flood plain and a point 7.2 miles eastward along U. S. 82, between Greenwood and Winona. Most of these are cuts through spurs sloping northward, across the road, toward the valley of Sand Creek. These spurs have been formed by dissection of depositional terraces, remnant flats of which occur at higher elevations in the immediate vicinity. Spur exposures exhibit pseudoanticlinal characteristics, with loess thickening down limbs. Contrasts between exposures on opposite sides of the road are evident. Transitions between terrace deposits and loess are shown on the first grade above the flood plain and 3.0 miles east of the flood-plain margin. At a gravel pit 2.1 miles farther east columnar loess contains admixed gravels and snail shells.

YAZOO CITY, MISSISSIPPI (Valley, Mississippi): Exposures essentially similar to those east of Greenwood occur for 11 miles eastward from the junction of 49E and 49W at Yazoo City along U. S. 49, toward Benton. More perfect gradational relationships may be seen near by on the road from Crupp to Anding, on the first grade rising from the flood plain. The great width of loess outcrops here is typical of central and southern Mississippi, where Montgomery terrace deposits are composed of backswamp materials and the surface lacks meander-belt drainage patterns.

VICKSBURG, MISSISSIPPI (Vicksburg, Mississippi-Louisiana): Excellent exposures occur along all roads in the vicinity of Vicksburg, the deepest cuts along newest highways offering the finest examples of gradations between loess and terrace deposits. Many excellent exposures in the National Park have been destroyed by artificial landscaping, but a ravine west of Wisconsin Circle exhibits gradational relationships, and beds beneath loess are exposed.

Eastward, toward Jackson, along U. S. 80, loess outcrops extend for about 20 miles, airline to Edwards. Exposures resemble those east of Greenwood (Pl. 3, fig. 1). Still more instructive exposures occur southward, toward Port Gibson, on U. S. 61. Loess has accumulated to considerable

depth in all valleys in this deeply dissected territory, and gradational relationships into terrace deposits occur only at higher elevations along the road.

Loess at Vicksburg appears unusually thick and continuous in outcrop. On U. S. 61, north of town, near Kings Crossing and Signal Post Office, it occurs close to flood-plain level. In the immediate vicinity it caps Fort Hill, one of the highest elevations along the bluff. The topography, however, is primarily the result of dissection in Tertiary materials. Excellent exposures of the Vicksburg group (Oligocene) occur along U. S. 61, and overlying beds on the road just north of the National Cemetery have been reported as Catahoula (Miocene). These Tertiary deposits extend well up the bluff, and loess is only a mantle.

NATCHEZ, MISSISSIPPI (Natchez, Mississippi): The classic localities in the vicinity of Natchez are probably the best in the lower Mississippi Valley. The first 3 miles on U. S. 84, east of Washington, toward Meadville, are particularly instructive. Backswamp clays dominate terrace deposits, and loess is sharply localized on slopes. Pseudoanticlinal exposures (Pl. 1, fig. 1) are numerous, and gravels may be traced to their outcrops. Eastward beyond the Adams County Line, no loess occurs along the entire road to Brookhaven. Topographic settings favorable to its development occur at many places, but east of Fenwick all terrace deposits are too coarse to serve as parent material.

South of Natchez, along U. S. 61, loess exposures are numerous to a point 10 miles south of Homochitto River. Pleistocene terraces and underlying Tertiary formations have been deeply dissected in much of this territory, and loess mantles lead down slopes, with increasing thickness, to the deepest valleys. Extreme range in elevation, disregard for compass orientation, and movement across truncated strata are especially well exhibited south of Homochitto River. The outcrops farthest south (Pl. 3, fig. 2) trace upward to Williana backswamp materials but nowhere into elevations above the Williana surface. Downslope they cross the coarser basal part of the Williana formation, where gravels become incorporated in the loess, the Bentley, and other formations. Other outcrops lead upward only as far as Bentley or Montgomery backswamp deposits.

SUMMARY

Field relationships demonstrate derivation of lower Mississippi Valley loess from backswamp deposits of Pleistocene formations, excepting that none is known to have been derived from the youngest, or Prairie, formation. It occurs on slopes leading up to fine-grained, calcareous, and organic parts of terrace deposits and is absent from slopes leading to coarser materials, such as characterize meander belts. It thickens downslope into valleys, forming a mantle generally conforming to the topography. It crosses truncated Tertiary and Quaternary beds and at many places incorporates materials from them, notably gravels from the lower parts of terrace formations, and, in some cases, lignite.

The thickest and most widespread deposits occur east of the Mississippi River, along bluffs and the sides of valleys in deeply dissected territory, especially in central and southern Mississippi. They are especially well developed where drainage patterns and sediments indicate the presence of Pleistocene backswamps. They are absent, or poorly developed, where drainage patterns indicate meander belts and coarse materials. Other conditions being equal, a deeply dissected territory favors development.

Restriction of loess to areas where the sediments have been deposited by Pleistocene equivalents of the Mississippi River may mean that only those rivers had flood plains broad enough to permit extensive backswamp deposition.

PHYSICAL CHARACTERISTICS OF LOWER MISSISSIPPI VALLEY LOESS

ESSENTIAL VS. NONESSENTIAL CHARACTERISTICS

An acceptable definition of loess must limit itself to essential physical qualities displayed by samples, so that laboratory identifications may be made as readily as for

FIGURE 1. LOESS CUTS

Typical exposures of loess on spurs, in cuts along U. S. 80, between Vicksburg, Mississippi, and the Big Black River, on the road to Jackson.

FIGURE 2. LOESS MANTLE

Loess overlying Bentley gravels in the high and deeply dissected hills of southwestern Mississippi, on U. S. 61, about 10 miles south of Homochitto River. This outcrop extends downslope across Tertiary strata on a slope about 200 feet to the right.

FIGURE 3. DELOESSIFICATION

The darker material toward the base of the exposure is loess, and lighter material, extending up to the base of the soil profile, is leached loess, on U. S. 61, 4 miles north of Sibley, between Natchez and Woodville, Mississippi.

FIELD RELATIONSHIPS

other types of rock. The definition discussed earlier confined itself to these essential features. In addition, certain other characteristics serve to differentiate it from other sedimentary deposits, but they may be regarded as nonessential.

NONESSENTIAL CHARACTERISTICS

COLOR: Ordinarily yellowish or buff, it may vary from ash gray to red. Generally exposures in Kentucky appear lighter than those farther south.

The color is in part related to chemical form and content of iron coatings on individual grains, as was pointed out by Richthofen (1882). To a greater degree it depends on calcium carbonate and the entire chemical composition of the deposit, as discussed by Ganssen (1922, p. 42–43).

The major resemblance between loess and loesslike materials is likely to be color. Udden (1898) noted that blue-gray clay on weathering resembles loess in color, structure, and the presence of calcareous nodules. Samples collected by the writer as being typical Mississippi River "blue clay" dried out in a few months in his office and so changed physically that many visitors unhesitatingly identified them as loess. They effervesced with dilute acid. Gow (1913, p. 299) cites a similar case in Iowa, where the original material was "blue clay" from a well. Such samples fail by a wide margin to meet grain size and sorting characteristics of loess however close they may resemble its color.

Reddish loess was reported by Lyell (1834, p. 118) from Wurtemberg in the vicinity of "*bunter sandstein*" outcrops. Veatch (1907, p. 298) found loess *Kindchen* in red loesslike materials along Red River in Louisiana. Reddish loess occurs east of Greenwood in the vicinity of seepages that keep it moist most of the time. In the lower Mississippi Valley reddish coloration appears to be associated with alteration, or pedologic degradation, into a type of brown loam.

VERTICAL JOINTING: This is characteristic of exposures near bluffs. Where best developed the fractures outline prismatic columns up to several feet in diameter and at places extend down into underlying materials. On the grade leading to the ferry landing at Columbus, Kentucky, the columns go down into gravels. They are present in Tertiary beds near the Herring Hill cut north of Memphis. Along sections exhibiting transitions into terrace deposits the perfection of jointing increases with calcareous content, being most strikingly developed in loess, and hence is related to greater competence associated with more complete cementation. Vertical jointing is wanting on flatter areas and best developed on steep slopes.

Field evidence indicates that vertical jointing is mechanically induced by tensional stresses associated with downslope creep and by settling. Todd (1897, p. 48) noted that many fractures widen upward and "might be compared to crevasses in a glacial rapid." Alternate wetting and dehydration may play some part in their formation. Roots may assist in forming vertical fractures, but their orientation is about vertical anywhere, and the jointing of loess is confined to slopes. Robinson (1936, p. 187) notes that in fine-grained materials fractures are likely to remain as permanent structural units over recurring cycles of moistening and desiccation.

VERTICAL EXPOSURES: Perpendicular faces occur principally in artificial cuts and where outcrops are undercut by streams. Their ability to remain almost unaltered

for years is related to cementation. In gradational changes between terrace deposits and loess, faces become more permanent and vertical as calcareous content increases. In artificial cuts where walls are not vertical, loess exposures have a smooth appearance. An abrupt contrast occurs where it is covered or underlain by other materials (Pl. 1, fig. 1). Leached loess weathers in a pinnacly manner, as do unaltered backswamp deposits. In some cases beds beneath the loess are firmly cemented and stand vertically in exposures.

Macfarlane (1884, p. 101) advanced the idea that vertical faces are the result of flat particles horizontally packed "like sheets of paper," but microscopic examination shows the absence of tabularity in typical constituent particles.

Vertical jointing and loess falls often determine natural faces but do not account for the ability of loess to retain them. Cementation is the main factor. It is present in primary form throughout the deposit, and concentrations of lime often form surface crusts. The latter give the effect of case hardening and render old cuts particularly durable, as was noted by Hilgard (1860, p. 314).

TUBULAR STRUCTURES: Swallow (1855, p. 70) noted these structures in loess, particularly toward upper surfaces. Todd (1878) observed that they reach maximum depths of 30 to 40 feet, being rare at more than 30. Pumpelly (1908, p. 7) reported a maximum depth of 100 feet, and Moyer (1932) 120 feet, which agreed with local depths of living plant roots. Practically all writers agree that the tubes are formed as coatings on roots (Todd, 1878; Broadhead, 1879; Howorth, 1882). Remains of roots are common within them in the lower Mississippi Valley and elsewhere (Todd, 1878; Baltz and Polynov, 1930). In German they are called *Wurzelröhren* (root tubes).

The smaller root tubes are less than 1 mm. long and constitute the greater part of the "sand fraction" when loess is subjected to mechanical grain-size analyses. A sample subjected to rapid washing on a 200-mesh sieve will have a much larger sand fraction than one broken down by methods involving longer soaking, agitation, and the introduction of some dispersing agent. This sand fraction, on microscopic examination, consists mainly of root tubes. Samples subjected to prolonged immersion in dilute hydrochloric acid exhibit few entire tubes, though frements of tube walls may remain. The cementing substance of the walls is mainly calcareous, but iron oxide may be present in various amounts, and some examples have been described as ferruginous by Safford (1869, p. 433) and McGee (1891, p. 437). Many have hollow cores. Opaque white lime fills some.

Root tubes may be simple, with openings at either end, or complex, with many openings of various sizes and with shapes similar to those of the more complicated loess concretions.

Willis (1907, p. 253) related root tubes to salts left by capillary waters. Richthofen (1882) attributed vertical faces to tube orientation, and others have expressed the same idea (W. B. Wright, 1914, p. 199). They are regarded by many as having some bearing on the origin of loess (Lobeck, 1939, p. 389). It has been contended that they mark positions of grass blades or shallow roots and that their vertical extent is proof that vegetation persisted during long periods of dust deposition. It seems reasonably certain, however, that they form at depth around living rootlets. Their

absence from brown loam and other loesslike materials may be explained by the smaller amounts of soluble salts, so that less is available for concentration around organic nucleii.

INCRUSTATIONS: Calcareous crusts are characteristic of fracture faces (McGee, 1891, p. 437) and exposed surfaces (Hilgard, 1860, p. 314). They form around exposed roots and below overhanging walls, some resembling stalactites. Any surface favorable to evaporation is likely to become encrusted. In the transition zone toward parent material the reduction in calcareous content is gradual. It extends throughout loess, but farther upslope only scattered blebs of the deposit effervesce with dilute acid. In typical parent materials no such evidence of incrustation appears.

CONCRETIONS: They are extremely abundant, variable in size, and irregular in shape. They have been given many names, such as *Lösskindel*, *Lössmanchen*, *Löss-puppen*, *Lösskinchen*, and *Stein-Ingwer*. Smaller concretions may form around root tubes, but organic nucleii appear to be absent from larger examples. Call (1882, p. 373) cracked open 2803 specimens without finding a single example, and the writer has had similar experiences. It is commonly stated that they are hollow, but the writer has never found an example resembling the large, rounded, hollow concretions so abundant in lower Mississippi Valley Tertiary and Quaternary gravel deposits that locally bear the name "rattle-stones." In his experience hollow concretions are ferruginous and bear little resemblance to the irregular calcareous concretions in loess.

Concretions are lost when loess is leached but are often found concentrated in great numbers immediately below leached layers. They are absent from parent materials and brown loam but grade into beds of limestone in gravels, sands, or other coarse materials underlying loess, particularly in bluff exposures.

HEALING OF FRACTURES: This is characteristic of loess, excepting that tensional gaps may be too wide to heal, and vertical joints may be so located that stresses tend to maintain them. Closed and compressed fractures often disappear soon after being formed. Todd (1899) was possibly the first to recognize this, attributing it to cementation during periods of dampness, noting that deposits that have suffered rearrangement closely resembled those in place. His observations may be verified throughout the lower Mississippi Valley.

An excellent example of healing may be seen 1 mile south of Natchez, on a grade leading to the flood plain at the U. S. Engineers Office yards. Toward the top of the grade a sharp line separates leached and unleached loess along the northern face of the cut. It bends with the topography and continues to a point somewhat more than half the distance to the foot of the grade, where it terminates abruptly, its former extension toward the river being displaced downward many feet. Lower in the section stratified sands and gravels dip back toward the bluff at about 20°, indicating rotation of a slump block. The termination of the line at the base of the leached loess farther upgrade marks the surface upon which rotation took place, but any fractures that were ever associated with the movement are no longer visible. So complete has been healing that no fracture remains though the aggregate throw has been more than 50 feet.

Healing of fractures may be noted in many cuts where loess underlies surfaces exhibiting terracettes or other evidences of displacement. Along highways loess falls

are relatively numerous on new roads but rare and obscure on old, where fractures between falls and loess in place have ordinarily healed. It is asserted that loess dwellings in China consisting in part of blocks of loess resting on floors of the same material become practically monolithic in a few years. Udden (1902) cites one of the few cases where shearing planes of considerable antiquity exist in loess.

GRAVELS: Gravels are common in loess. Lyell (1834, p. 119) observed them in practically all exposures he examined. Carmony (1903) emphasizes their abundance. Call (1889) related them particularly to lower parts of exposures along Crowleys Ridge. Similar observations have been made by Mabry (1898), Glenn (1906), Chamberlin and Salisbury (1907), and others. Carman (1929) associated them with gravel deposits nearby, and this is the key to their presence.

Gravels are genetically foreign to loess but become incorporated during mass movement. The higher part of an exposure is ordinarily free of gravel, as are the topographically higher parts of outcrops. Gravels may in all cases be traced upslope to their source beds. Loess creeping over gravel layers (Pl. 1, fig. 1) picks up stones and incorporates them in process of carrying them downslope. It also picks up sand, fragments of Tertiary beds, or other material that it crosses. For this reason a mechanical analysis of samples collected low in outcrops has little meaning from the standpoint of essential loessial characteristics.

Gravels have seemed incompatible with eolian origin to many persons. We encounter such statements as, "If it occurred without the gravel it would at once be pronounced loess," with reference to a deposit in central Asia (Huntington, 1905, p. 178). Gravels have been used as an argument in favor of fluvial origin, but most examples appear to be loesslike materials. Barbour (1928) regards them as evidence of reworking. Where reworking involves fluvial action and loess is carried away to be redeposited the gravel will be stratified and even the finer parts of the deposit will no longer resemble loess in any essential characteristic. The random orientation of stones in the case of gravels in loess is evidence of the mass movements that permitted its incorporation.

TOPOGRAPHIC FEATURES: Many are bizarre, such as loess wells, bridges, pipes, deep ravines, and high, narrow divides. Pumpelly (1866, p. 40; 1918, p. 468) popularized them. Fuller (1922) gives excellent descriptions, and Landes (1933) adds much, especially with regard to caverns. One of the best general statements is by Scheidig (1934, p. 26–41).

Underlying beds often have much to do with loess topography. Unconsolidated layers that are removed by sapping often determine more or less continuous lines of caves immediately below exposures. Sapping and progressive piping in underlying beds favor subsidence of cavern roofs. Vertical fractures favor subsidence producing terracettes that may resemble long flights of stairs. It is cementation of loess and relative competence that differentiates the topography from that of near-by areas.

THE LOESS FAUNA

Terrestrial snails form an overwhelming element in the fauna. Most authorities agree that all species are now living. Lyell (1834, p. 120; 1838, p. 84) was first to

observe this and first to state that every species now inhabits the immediate vicinity (1849, p. 150, 156). McGee (1891, p. 471) found this to be the case at Davenport and Muscatine, Iowa. Upham (1895, p. 283), E. H. Barbour (1903), Hay (1914, p. 42), and many others have noted the same thing. Shimek (1896; 1902) emphasizes the point by such statements as: "Every species of molluscs which has thus been reported from Iowa, Nebraska, and Missouri is living today, and with three or four exceptions all are found living within the territory covered by the loess" (1897, p. 33).

The similarity between living snails and those in loess is amazing in the light of the fact that the group is so highly differentiated and so many species are sharply restricted as to locality. Its significance is apparent in the fact that loess grades upslope into parent materials ordinarily devoid of fossils. The snail shells, of necessity, are incorporated sometime during, or after, loessification.

That snail shells favor the hypothesis of eolian origin has been advocated by many, notably Shimek (1897), who correctly emphasized the overwhelming predominance of pulmonate types. His arguments were directed mainly against the lacustrine hypothesis and from that standpoint were sound. A flora rich in land plants must have been available to these herbivorous creatures (Shimek, 1897, p. 39). Lyell, on the other hand, noted that snails on the flood plain of the Rhine were the same types as found in near-by loess (1838, p. 84) and that those alive on the shores of Lake Concordia, Louisiana, were identical with those in loess at near-by Natchez (1849, p. 156), and from such facts argued for fluvial origin. Winchell (1903, p. 145) noted that snails creeping out on flood plains would become incorporated in alluvial deposits, and this happens commonly in the lower Mississippi Valley, examples having been found on the campus of Louisiana State University, at Giles Bend Cut-Off, near Natchez, and at a great many other places. Their absence from alluvial deposits that now form terraces is explained by leaching.

The fact that snail shells are so common in loess and so rare in loesslike materials does not mean that the calcareous content of the former has been derived from the shells. Specimens are not corroded or in process of dissolution. As a rule they are entire, and many retain original color markings. The amount of lime in loess is many times greater than the amount available from shells.

Terrestrial snails thrive on limy soils. Hesse *et al.* (1937, p. 416) show that larger and heavier shells, in proportion to body weight, characterize individuals growing where lime is abundant. In modern flood plains snails are far more abundant in calcareous backswamps than on less calcareous natural levees. Snails both living and dead are found in calcareous environments. They seek the type of place where loess is forming and in incorporation encounter the possibility that their tests will be preserved.

Kingsmill (1871) was possibly the first to observe that snails in loess are not fossils in the ordinary sense of the term. He called them "pseudo-fossils" and attributed their presence to falling in fissures. Some may be present for that reason. In places they occur in pockets that suggest intrusion. Others may have burrowed into the loess. Exposures along new road cuts, however, demonstrate widespread occurrence, individual separation and scattering, lack of definite orientation, and thus suggest that the typical mode of incorporation is during mass movement. Those

finding their way into loess are preserved in a calcareous environment, but those getting into other materials are subjected to leaching and disappear promptly.

From the stratigraphic standpoint the snails can indicate nothing more than Recent or late Quaternary age. They might be used to date the time of loessification but not the age of the parent material. In leached loess all traces of shells are lost.

In the first monographic treatment of loess faunas Braun (1842) established a custom that has been followed by many investigators, that of counting numbers of individuals of each species and variety identified. In the Rhine Valley, between Basel and Bonn, 211,958 specimens, regarded as representing 30 species, were distributed in the following proportions among genera: *Succinea*, 47%; *Helix*, 39%; *Pupa*, 13%; and *Clausilia*, 1%. The genera *Bulimus*, *Limneus*, *Limax*, *Planorbis*, and *Vitrina* occurred in small numbers at many localities. *Succinea oblonga* was the most abundant species along the Rhine, and also near Krakau, Poland, along the Danube, near Linz and Pressburg, and along the Rhone, near Lyon. In southern France Braun found *Helix hispida* the most abundant species, but the writer doubts that the specimens came from loess in that region. Of his over 200,000 specimens Braun found only 33 that were strictly water dwellers. While Shimek and others have expanded faunal lists greatly and have changed various generic and specific names, their main conclusions agree with those of Braun; the shells in loess are those of land animals now living in the vicinity.

Next in abundance to terrestrial snails in loess are fresh-water ones. They are relatively rare, and those reported come from a comparatively small number of localities. They probably lived in small ponds such as now exist, here and there, in the vicinity of loess deposits. Other mollusks, such as Unios, reported from loess, probably come from loesslike terrace deposits or else have been incorporated along with gravels and other extraneous materials.

Vertebrates have been reported from loess at many places. Lyell (1838b; 1838c) found a shark in Rhine Valley loess. Aughey (1880) lists teeth, fish bones, rabbits, gophers, otters, beavers, squirrels, deer, elk, buffalo, mastodons, and other mammals from loess encountered in wells near Lincoln, Nebraska. E. H. Barbour (1903) reported similar finds. Harris and Veatch (1899, p. 116) refer to a mastodon found in the basal part of the loess. Many mastodons have been so reported from the vinicity of Natchez and southward into Louisiana. Hilgard (1879, p. 110) commented on the scattered condition of bones in these finds. In collecting vertebrate fossils in this territory the writer has found them only in terrace deposits. Hay (1914, p. 42) regards the occurrences of vertebrates in loess as "mainly questionable." While there is no reason why a tooth or bone should not find its way into this material, just as gravel does, one may suspect that such important fossil deposits as those in the vicinity of Lincoln, Nebraska, actually were in terrace deposits. Vertebrates and aquatic mollusks do not appear to be significant parts of the lower Mississippi Valley loess fauna. If this happens to be true in Europe a great many ideas about climatic conditions under which the deposits formed are subject to change. The writer is familiar with the Heidelberg Man locality at Mauer. The vertebrate finds

were near the basal part of a depositional terrace sequence, in coarse deposits, and loess lies on slopes above, apparently originating from finer beds, such as characterize the upper parts of many terrace formations.

ORIGIN OF LOWER MISSISSIPPI VALLEY LOESS

GENERAL STATEMENT

No theory in vogue today appears to account for the physical characteristics, distribution, or stratigraphic position of lower Mississippi Valley loess. The closest approach has been advanced by Berg (1932), but insistence that it be formed under arid climatic conditions is locally untenable.

The essential part of loessification, or the process by which loess is differentiated from its parent material, is concerned with producing from backswamp deposits a sedimentary material having the typical grain size, sorting, porosity, and calcareous content of loess. No widely held theory seems capable of doing this.

OBJECTIONS TO EARLIER LOESS THEORIES

The objection may be raised against fluvial or lacustrine origin on the basis that loess is not a stratified deposit *in situ*. The Pleistocene deposits of the lower Mississippi Valley are fluvial terrace formations, and their stratigraphy is thoroughly understood. Loess forms no part in the stratigraphic sequence. It extends down slopes across truncated edges of both terrace deposits and Tertiary strata. In vertical range it lies between the higher, finer deposits of the oldest, Williana, terrace and the active flood plain. In southwestern Mississippi at many places this range amounts to over 300 feet within 1 or 2 miles. Similar observations have been made elsewhere.

Lyell (1834, p. 120) found it overlying all other formations in its vicinity, including the latest gravels. In Indiana, Shaw (1915, p. 107) gives the vertical range of "bluff loess" as from 700–800 feet at its top to 375–400 feet at its base. Worthen (1866, p. 58–83) found that in Illinois it occurs as coatings on bluffs, failing to reach their tops but extending to their bases, as did Fowke (1908, p. 37) near St. Louis, Missouri. Richthofen (1877, p. 156) reported deposits near Vienna that extend upslope for 300 feet, nowhere more than 50 to 60 feet thick at any given place. Barbour (1929, p. 64) stated that in China: "Though loess can be traced uphill uninterruptedly for distances of more than 500 feet, the greatest vertical thickness actually measured in unquestionable primary loess is only 35 feet." Such observations preclude a lacustrine, fluvial, or marine origin.

Against eolian origin it may be urged that no actual or hypothetical directions of winds could account for its distribution. It covers slopes leading in all possible directions and is ordinarily as strikingly developed on one side of a ridge as on the other. In pseudoanticlinal exposures, whatever their orientation may be, one limb ordinarily resembles closely the other. It occurs on both sides of the Mississippi and other large rivers.

The shape of a deposit is that ordinarily assumed by colluvial materials. Richthofen (1877, p. 154) described it as "leaning" against mountains and (1882) likened its upper surface to a rope stretched loosely between two ranges, a fortunate simile that has been quoted many times. Such shapes are typical of lower Missis-

sippi Valley deposits. Nowhere are such forms as dunes or deposits heaped with steep faces toward the valley and gentle slopes leading eastward to a feather edge as many supporters of the eolian hypothesis believe.

The sorting appears to be too uniform to be the result of direct deposition from a current. It seems improbable that either wind or water could move with the uniform velocity required to permit the accumulation of material so homogeneous that at least a half and ordinarily about three fourths of its particles (by weight) fall within the limited diameter range 0.01–0.05 mm.

The ratio between calcium and aluminum is one not likely to be encountered in any variety of primary sediments (Polynov, 1937, p. 175). Detailed studies by Krokos (1926) show that Ukranian loess could not be a product of mechanical disintegration of rock, such as rock flour or dust.

In the lower Mississippi Valley loess development correlates mainly with two main factors: (1) the presence of backswamp deposits in terrace formations, and (2) deep dissection. To be acceptable a theory of origin must harmonize with these facts.

LOESSIFICATION

The process of loessification starts in parent material that originally was deposited as alluvium on flood plains during the Pleistocene. It affects the finer parts of such deposits, especially those that have accumulated in backswamps and are present only in minor amounts along Pleistocene meander belts. It is restricted to parts of terrace formations that now stand considerably above flood plains. The deposits must consist mainly of silt and clay. They are somewhat calcareous and contain carbonaceous matter derived from plant remains.

The initial stage of the process is weathering and differentiation of soil profiles. While pedogenic processes are active much of the original calcareous content, including any fossil shells that may be present, is lost to ground waters. The resulting product is a brown loam that thickens residually on flats but is relatively mobile on slopes. In deeply dissected territory it creeps into valleys, where it accumulates to considerable thicknesses.

The colluvial phase of the brown loam is derived from the upper parts of the profile of weathering and soil development and hence is characterized by coarser particles than the average present elsewhere. The loss of finer materials goes on at all stages of colluviation and is intensified by churning movements. Surface washing probably contributes to some degree. With increasing distance downslope comes closer approach to the remarkable sorting and uniform texture of loess.

Toward the lower parts of colluvial slopes is a zone of carbonate enrichment, the carbonates having been derived from terrace materials and brown loams on surfaces upslope. Snail shells introduced during colluviation are preserved only where carbonate enrichment takes place and hence characterize materials advanced far in loessification. The introduction of carbonates effects a measure of structural competence, retards creep, makes fracturing possible, and renders faces relatively stable. By the time significant enrichment has occurred loessification is practically complete.

There are several distinct processes involved in loessification. We may turn to their details.

CALCAREOUS CONTENT: Mature loess contains 10 to 25% calcareous content (Keilhack, 1920, p. 157), varies from 3.25 to 25.60% within vertical sections of Ukranian loess (Krokos, 1926, p. 117), or, on the basis of weight lost under hydrochloric acid treatment, is somewhat less than 36% for lower Mississippi Valley deposits. Concentration of salts is one of the most essential parts of loessification.

Bluffs facing flood plains are especially favorably located from the standpoint of salt accumulation. Low positions in valleys of deeply dissected territory are also favored. Substances leached from topographically higher, calcareous backswamp deposits are carried by ground water to lower levels, and many are concentrated toward exposed surfaces.

The calcareous content of loess is secondary according to most observers. Chamberlin and Salisbury (1885, p. 304) emphasize the point that the amount present in snail shells is disproportionately small and that a magnesium carbonate element could not have come from shells. Willis (1907, p. 249) regarded water as the depositing agent. Hay (1914, p. 41), Krokos (1926), and others have noted increasing calcareous content toward lower positions in exposures.

Polynov (1937, p. 175) notes that alkalies derived from the leaching of such materials as backswamp clays move in ground water mainly as bicarbonates. During capillary movements in which the solutions approach the surface an intensive separation of CO_2 takes place, and the remaining salt is precipitated as the normal carbonate.

"We can thus assume that the calcareous south Russian loess is a region of concentration of that lime (and partly magnesia) which was lost in solution some time by the moranic drifts adjoining the loess on the north, and now to a considerable extent depleted of their alkali and alkaline earth bases, and in particular calcium. Analogous relations may be observed in over more limited areas in Northern Mongolia, where the leached crust of weathering covering summits and upper slopes of the Kentei and Khangai massifs passes at lower levels into a mantle of calcareous loess."

The source of the carbonates in lower Mississippi Valley deposits is evident for they occur below the most calcareous parts of Pleistocene flood plains. Pleistocene meander-belt deposits are not only unfavorable to loessification from the physical standpoint but also because they are less calcareous. The transitional increase in calcareous content of materials involved in downslope gradations between terrace deposits and loess and the striking concentration of salts at low levels is proof that accumulation is of secondary origin.

Materials other than loess collect lime under similar conditions of exposure in the lower Mississippi Valley. Bluffs of sand and gravel are more calcareous than their stratigraphic equivalents in flatter territory. Huge volumes of concretionary material are common in such deposits, and along some layers tabular masses of secondary limestone have accumulated. Excellent examples occur south of the bridge at Natchez.

Lime cements many of the grains in loess into coarse aggregates. Surplus lime forms concretions. A surplus is ordinarily available, so that even rearranged deposits develop concretions (Todd, 1899, p. 99). Exposures favorable to the escape of moisture develop crusts, as was noted by Hilgard (1860, p. 314), Witter (1892b, p. 330), and many others.

The mineralogical investigation of carbonates such as occur in loess is practically a virgin field for research according to Polynov (1937, p. 31–32).

GRAIN-SIZE CHARACTERISTICS: In typical backswamp deposits there is less sorting and, on the whole, much finer material. Loessification witnesses a change in which sorting becomes emphasized and the size of the average particle is increased.

The idea that loess grades into alluvial clays is not new. Hilgard (1879, p. 107) noted loess passing laterally into clayey loam. Leverett (1897, p. 157) records loess passing into clayey loam and then into loamy clay. Mabry (1898) emphasized the close relationship to brown loam. Harris and Veatch (1899, p. 117) regarded loess as a special development of loam. Todd (1894) noted that loess contains more clay away from streams until it so resembles adjoining clays that it is difficult to separate the two. Shaw (1915, p. 107) also found increasing clay content with greater distance from streams. This observation has been used to support eolian origin,—it being supposed that coarser particles lodge lower on bluffs. Some believe the entire deposit was loess originally, the clays and loams being an expression of weathering (Mabry, 1898; Sardeson, 1899).

Earlier works support the idea that loess is colluvial. Hilgard (1860, p. 195) noted that the deposits thicken downslope, and Moyer (1932) found the same true in China, as did Worthen (1868, p. 123) and Winchell (1879, p. 168) in the central United States. Chamberlin (1890, p. 471) and Todd (1918, p. 116) found upper and lower parts of exposures connected as the result of creep.

A relationship between colluviation and uniformity in grain size was suggested by Worthen (1866, p. 329), who visualized a "partial sifting process" as taking place on slopes. Wahnschaffe (1886, p. 360) thought that some German loess is still being formed as the result of removal of finer materials while it occupies positions on slopes. That loess has been derived from deposits including a larger proportion of finer materials is demonstrated by mechanical analyses, such as those of Table 1, for these invariably show a tail of fine grain sizes extending far beyond the coarse silts that constitute an overwhelming proportion of each sample.

There are more factors in the "partial sifting process" than Worthen, Wahnschaffe, or others contemplated. It is initiated during weathering, intensified by pedogenic processes, stimulated during colluvial transportation, and appears to end only after the deposit has become thoroughly cemented. Selective truncation and concentration of upper, coarser parts of the profile is characteristic during each stage of loessification. Downward translocation of finer particles occurs during each stage. Other losses, such as material entrained by various currents at the surface, favor removal of finer and concentration of coarser particles in the residual material.

The transition between parent material and loess is so gradual that field geologists in the lower Mississippi Valley encounter difficulties in mapping contacts between lithologic units. Many Tertiary and Quaternary deposits yield a brown loam that closely resembles loess. As a group the brown loams are relatively noncalcareous and do not effervesce with dilute acid. Their loss in weight when treated with hydrochloric acid in the laboratory is ordinarily about 5 per cent, and this includes removed free iron oxides and soluble colloids as well as carbonates. Mechanical analysis ordinarily indicates degrees of sorting and materials of sizes quite unlike

those of loess. Brown loams taking part in loessification attain such close resemblance that mechanical analyses of samples collected near loess are almost identical, and calcareous content is the best guide as to whether it is loess or not. The samples in Table 2 indicate the last step in loessification from the grain-size standpoint.

The comparison presented in Table 2 is typical of that between loess and material immediately upslope that has undergone all stages but the final one, lime enrichment, in loessification. At the outcrop the material of sample 6 effervesced, contained snails, and was covered by moss requiring a calcareous environment, whereas all these features were absent in the other case. The loess contained somewhat more

TABLE 2.—*Comparison between loess and nearly loessified brown loam*

(In per cent)

Per cent of sample (by weight) finer than	Loess		Loesslike brown loam	
	6	6D	7	7D
Mm.				
0.06	94	97	99	99
0.05	86	90	97	98
0.04	76	82	94	96
0.03	63	71	84	90
0.02	44	54	52	66
0.01	18	24	20	32
0.009	14	21	18	28
0.008	11	19	16	26
0.007	11	18	15	24
0.006	10	17	14	22
0.005	9	17	13	21
0.004	9	16	12	19
0.003	5	13	10	17
0.002	0	11	6	15
0.001	0	8	4	13

Description of samples: 6. Loess from cut shown in Pl. 1, fig. 1, 1.6 miles east on U. S. 84 from junction with 61 at Washington, Miss. Sample taken at central part of exposure. Of the original material 19.7% was soluble in dilute hydrochloric acid. Analysis run without dispersing agent, with slight flocculation toward end of analysis.

6D. Same sample, same type of analysis, but with sodium oxalate dispersion.

7. Brown loam from next cut east, 0.4 mile away, at slightly higher elevation. At exposure was loesslike in all regards except that it would not effervesce with acid. In laboratory 5.2% was soluble in dilute acid. Analysis run without dispersing agent and without flocculation.

7D. Same sample, same type of analysis, but with sodium oxalate dispersion.

coarse material, 37% of sample 6 and but 16% of sample 7 being material larger than 0.03 mm., before dispersion. After treatment with sodium oxalate the proportions were, respectively, 29% and 10%.

In all cases in Table 2 dispersion reduced the amount of coarse material present, indicating the presence of aggregates that were to some extent broken down into constituent particles.

Berg (1927; 1932) regarded aggregation of clay particles into granules of silt size as the dominant factor in producing the sorting characteristics of loess. His evidence rests heavily on an experiment by Ganssen (1922, p. 41), who subjected a sample of

kaolin to the action of alkaline silicates (Table 3). Loesslike texture was produced in kaolin. Ganssen gives few details concerning his methods. The end product was certainly not loess. The experiment was conducted as an attempt to explain how wind-transported dust might be changed to resemble loess. Ganssen apparently failed to realize that his experiment might be used to discredit the eolian hypothesis, and Berg applied his results to that end.

TABLE 3.—*Textural change in a kaolin sample*
Subjected to exposure in a solution of alkaline silicates

Grain size of particles	Original material	After treatment
above 0.05 mm.	3.3%	11.6% (weight)
0.05–0.01	3.3	43.2
below 0.01	93.5	45.3

The lower Mississippi Valley affords ideal conditions for testing Berg's hypothesis. Pleistocene backswamp deposits, to a peculiar degree, contain all the elements favorable to consummation of the process of building aggregates out of finer particles.

Clay suspensions are flocculated by calcium salts (Baver, 1940) such as exist in backswamp deposits. Flocculation itself, however, does not insure the stability of aggregates. Removal of the flocculating agent may result in dispersion to original physical state. Where aggregates are cemented they may become granules that retain identities even after being exposed to an environment unfavorable to aggregation. Backswamp deposits not only contain the necessary alkaline materials to favor aggregation but also several agents favorable to preservation of granules.

The relatively high organic content of backswamp deposits favors stability of aggregates. Improvement in physical properties of clay soils treated with lime and humus is chiefly the result of aggregation of smaller particles into larger, relatively stable units. Lime promotes aggregation and with humus produces relative stability. Some contend that humus alone accomplishes both ends (Myers, 1937). That carbon deposits are disseminated in loess has been noted by Hershey (1896, p. 296), Udden (1898), and Shaw (1914, p. 298). They are common in residues of coarse fractions from mechanical analyses of lower Mississippi Valley loess.

Aggregate stability is also favored by the irreversibility of colloidal iron hydroxide with respect to dehydration effects. In many soils this factor is important in producing stable aggregates. Roberts (1933) has noted aggregation of 95 per cent of all original clay and silt particles into units larger than silt in the iron-rich Nipe clay of Puerto Rico. Lower Mississippi Valley backswamp deposits and loess are both relatively rich in iron.

Cemented aggregates are characteristic of loess. Hilgard (1860, p. 110) noted this and stated that they are permanently fixed by calcareous incrustations. Todd (1899, p. 95) found coatings of both calcium carbonate and iron oxides. Various European investigators (Scheidig, 1934, p. 61) regard *Krümelstruktur* (crumb structure) as characteristic of loess.

The degree of aggregation in a sample depends largely on the treatment it has received prior to examination. In thin sections the whole body is cemented into es-

sentially a single mass. Samples kept in bags form lumps that are aggregates in much the same sense that clods are in soils. Powders on slides exhibit clusters that are essentially aggregates of dimensions that depend chiefly on the amount of crushing they have experienced. Sand and silt fractions retained after vigorous treatments in mechanical analyses still display aggregates. The fundamental size of particles in such aggregates is that of silt, rather than the clay of Berg's hypothesis. Individual minerals ordinarily appear fresh and angular.

Tables or curves of grain sizes in loess attempt to refer to individual mineral particles rather than to aggregates. Their validity depends to a great extent on the treatment of the sample in analysis. Samples washed gently on 200-mesh sieves display exaggerated sand fractions. Violent agitation, long soaking, and effective dispersion must occur in order that most aggregates be broken down. In samples 6D and 7D of Table 2, resolution to fundamental mineral sizes is about as complete as possible. Few residual aggregates were found in petrographic examination of the various fractions, even in the case of the finest, when examined for Brownian movement. Such aggregates as were found consisted of material mainly larger than clay-particle size.

The most rigorous method followed in the present study involved treatment with dilute hydrochloric acid at various temperatures below boiling for prolonged periods, until every sign of effervescence stopped. Samples were then washed, the water being drained through a Pasteur-Chamberland filter. The loss in weight varied from between 19.7 and 35.8 per cent in loess and between 5.2 and 5.8 per cent in loesslike materials. Mechanical analyses were by pipette method, with sodium oxalate as the dispersing agent. No sample thus treated showed appreciably more clay than samples subjected to ordinary mechanical analyses. Petrographic examination of dried residues of various size fractions showed that loess is composed mainly of primary mineral and rock particles in such proportions as are cited in Tables 1 and 2. The silt particles are not aggregates of clay-size material, as postulated by Berg. Parent materials, well advanced in loessification, differ mainly with regard to carbonate enrichment, not with regard to constituent particles. Those less advanced differ mainly in having greater proportions of fine material. In all cases quartz grains are predominant. Calcite grains are conspicuous. Minor amounts of rock fragments, various heavy minerals, and other minerals characteristic of flood-plain deposits are present in varying proportions.

Porosity: This is an essential feature of loess. Most writers note conspicuous openings and many comment on the ease with which water travels through them. These descriptions are concerned mainly with visual porosity. Examination under a hand lens reveals open ends of root tubes, irregular fractures, crumb structures, and a general appearance not unlike moist sand. Specific porosity is comparatively low, much lower than that in backswamp deposits, but individual void sizes are large, so that loess is highly permeable. The high coefficient of permeability promotes effective drainage, encourages leaching at higher levels and soluble salt enrichment below, lowers water tables, and favors deep penetration of plant roots. All such features give loess a characteristic appearance that differentiates it from brown loam or other loesslike materials.

Todd (1897), Leverett (1897, p. 156), and others have noted that permeability decreases away from bluffs where the material exhibits less complete loessification, finer grain sizes, less structural competence, and fewer root tubes either on slopes above typical deposits or at depth behind exposed faces.

Berg (1932, p. 141) regards porosity as the result of aggregation of clay particles into coarser units, so that voids are left as a result. There is not enough fine material present during the final stage of loessification to make this possible, and its absence from loess aggregates disproves the idea. Aggregates affecting visual porosity are common but are composed of coarse materials.

RELATIVE COMPETENCE: This character of loess as contrasted with loesslike materials is the result of cementation. The colluvial transportation involved in loessification is inhibited or prevented so that loess undergoes only such mass movements as are characteristic of rocks of some strength and durability. Rock fall of a variety that may be called loess fall and slump are typical. Open fractures may be maintained. Such competence disappears gradually when loess is leached, and it is ordinarily wanting in parent materials or other loesslike deposits.

SUMMARY: Loessification is the process through which loess is formed from its parent materials. In the lower Mississippi Valley the initial parent material is Pleistocene backswamp sediment. Weathering and pedogenic processes convert its upper part into brown loam that is relatively noncalcareous, is devoid of snails, and tends to display pinnacly surfaces on steep exposures. It creeps freely on slopes and undergoes textural changes, mainly in the direction of losing its finer particles, in the process. Residual concentration of coarser materials, which are mainly of silt size in the initial parent material, starts with the weathering and pedogenic phase of loessification and continues until the final stage, when mass movement is checked by cementation. Vertical translocation of smaller solids is essentially equivalent to that taking place in the differentiation of the A-horizon of a soil. Surface losses of fine material occur for various reasons. Material involved in creep is that present at the surface and therefore the coarsest in typical profiles. There is thus a progressive loss in proportion of finer particles downslope and a consequent approach to grain sizes and sorting characteristics of loess in that direction. Carbonates leached from upper surfaces are translocated in solution and concentrated toward lower parts of slopes, where loessification becomes complete. Snails living on slopes are probably incorporated in moving materials at various places, but their shells are preserved only where carbonate content is increasing and for that reason occur only in loess.

AGE OF THE LOESS

Writers ordinarily agree that loess is Quaternary in age. Some restrict it to the Recent (G. F. Wright, 1901), citing such evidence as the finding of Indian arrow points at a depth of 12 feet at Muscatine, Iowa (Witter, 1892, p. 277). European literature ascribes interglacial age as a rule. In the Central States it is said to be interbedded with moraines.

There are really two values involved: (1) the age of the parent material, and (2) the date of loessification. In the lower Mississippi Valley the oldest parent material is early Pleistocene, Williana formation. The most widespread deposits are derived

from comparatively late Pleistocene, Montgomery formation. None appears to be derived from latest Pleistocene, Prairie formation, nor from the Recent. All the parent materials are interglacial in age, but this does not mean that loessification was also interglacial.

In Mississippi are widespread blankets of brown loam derived from Tertiary deposits, and many have suffered extensive colluvial creep, but no case is known where loess has been produced from it. The reason has not been sufficiently studied, but in some way Tertiary-derived brown loam lacks essential ingredients, possibly insufficient lime or the right proportion of silt. More enigmatic is the absence of loess derived from Prairie brown loam. Backswamp deposits are widespread, and many of the deposits are extremely loesslike. The reason may be the limited distance available for colluvial transportation, as the ordinary vertical interval between the Prairie terrace and the flood plain is 50 feet or less. During the maximum dissection of the pre-Recent glacial stage, however, maximum relief was increased by over 200 feet along the Mississippi Valley as far north as Cairo. Possibly not enough time has elapsed since Prairie deposition for complete loessification, or the process may require such environmental changes as accompany more than one complete cycle of changes from interglacial, through glacial, and back to interglacial conditions.

Europeans generally favor the idea that loess formed under arid or semiarid climatic conditions. The argument hinges largely on the presence of faunas in western Europe that now characterize steppes to the east. It has also been advanced from a more theoretical standpoint by Berg (1927; 1932), Rungaldier (1933), and others, who consider that carbonate enrichment took place under conditions resembling those that now form caliche crusts along valley sides in the Southwest. Such may have been the conditions in Europe, but it is wholly unreasonable to suppose that they ever existed in the lower Mississippi Valley.

Pleistocene North America had about the same essential outlines as the continent has today. Its relief may have varied somewhat but not enough to upset greatly the distribution of its air masses. During glacial stages the Polar-Canadian air mass was modified considerably by the presence of ice sheets, and during interglacial stages it more nearly approached its present condition. The Tropical-Gulf air mass to the south suffered less modification, remaining warm and moist at all times. The lower Mississippi Valley must have remained a major theater of interaction between these contrasted air masses. Storminess may have increased considerably during glacial stages, when their differences were sharpest, and precipitation may have been notably heavier and far more intense than it is today. Nothing short of a high transverse mountain barrier could keep these air masses sufficiently separated to allow even semiarid climates in the valley below Cairo. Available paleobotanical evidence (Brown, 1938) contradicts the aridity thesis.

A distinction has been made between arid atmospheric and arid soil climate by Ganssen (1922), who insists that the latter is requisite for carbonate accumulation in loess. Lowered water table favors soil climate aridity. Some lowering of water tables occurred during each glacial stage, when water in the Gulf of Mexico stood low and the main valleys were correspondingly dissected. As the depth of Recent alluvium is now 160 feet at Cape Girardeau, Missouri, 230 feet at the Louisiana-

Arkansas line, and 260 feet west of the bluff at Natchez, the relief immediately prior to the start of the Recent was locally increased over its present amounts by such values. When Recent deposition started, the bluffs at Natchez were over 450 feet high, and subsurface borings show that the face now covered by alluvium was about as steep as the part exposed today. This may be an important factor in the localization of loess deposits along bluffs facing the Mississippi and the valleys of its principal tributaries. A more arid soil climate along such bluffs may have favored carbonate accumulation. It seems probable that the longer colluvial slopes themselves had more to do with the presence of loess and that carbonate accumulation occurred toward the bases of slopes without aridity either in atmospheric or soil climate.

In summary, the age of parent materials is a simple matter that has been determined, but the date of loessification is more involved and cannot be stated precisely. There may have been several loessification stages. Each may have been a glacial stage, when valleys were more deeply dissected and colluvial slopes were longer. The correlation between loess distribution and deeply dissected topography is suggestive in this regard. There may have been a time factor as well, so that only longer glacial stages were effective. The fact that Prairie deposits nowhere serve as parent materials is suggestive. The pre-alluvial topography of the lower Mississippi Valley indicates a comparatively short pre-Recent glacial stage. The Mississippi, Ohio, and other rivers ran through comparatively narrow and steep-walled valleys in a region of low, rolling hills immediately prior to the deposition of basal gravels of the Recent alluvial sequence. Time was insufficient during the glacial stage for much lateral corrasion or the development of valley flats. On the other hand, the absence of loess in Prairie and Recent colluvium may be related only to the time factor. Loessification may be in progress today, and its rate may have been quite uniform for a large part of the Quaternary, but completion of the process may require an interval that has affected only pre-Prairie parent materials as yet.

DELOESSIFICATION

The loss of essential characteristics by loess may be called deloessification. It is promoted by conditions favorable to leaching, but the entire process involves far more than carbonate removal. It is particularly active in the upper parts of artificial cuts.

The lower limit of deloessification is ordinarily a sharp line between loess, below, and altered material above. In Figure 3 of Plate 3 this line appears about one third of the distance from the bottom of the cut in the central part of the exposure and slopes downward on either side. Below it is calcareous material, with concretions and snails, darker in color, especially when damp. On the loess surface is a cover of moss. The material above lacks all these evidences of lime.

While the "non-calcareous loess" of literature may mean either deloessified deposits or loesslike materials, a sharp distinction should be made between the two. Loesslike materials have never been loess, and most of them never will be. Deloessified material retains many loessial characteristics. It is practically the same from the standpoint of particle sizes and sorting, as shown in Table 1. Sample 4 was collected 6 inches below the bottom of the zone of deloessification, and sample 2, 6 inches above,

the materials having been only 1 foot apart at the exposure. An increase of 4 per cent in fine material in the range 0.001–0.007 mm. and decrease of 5 per cent in material coarser than 0.05 mm. suggests a slight breaking down of aggregates during deloessification.

The initial expressions of deloessification occur in an amazingly short time. There is a direct correlation between depth of the altered zone and the number of years of exposure. The newer cuts along U. S. 61 south of Vicksburg now display deloessification zones many feet thinner than those along old roads in the vicinity. The zone is always thickest on the face of a cut, and ordinarily its base slopes sharply upward in the first few feet.

In pseudoanticlinal exposures concentric fractures often exist in the comparatively competent deloessified parts of profiles, but they are not retained by material that has again become brown loam.

The base of the deloessified zone follows topography about as water tables do. In pseudoanticlinal exposures on spurs the altered material is thinnest at the crest and thickens down either limb. In an excellent example (Pl. 1, fig. 1) its thickness is about the same as the loess below.

Effective leaching removes carbonates, shells dissolve completely, root tubes suffer various degress of destruction, and aggregates start breaking down in the early stage of deloessification. Moss, regarded by Hilgard (1884, p. 302) as largely responsible for smooth faces of loess exposures, no longer thrives on the surface. This, the disappearance of calcareous crusts, and decreasingly effective cementation permit rill development, so that pinnacly surfaces replace the smooth faces characteristic of loess outcrops. All features related to competence gradually disappear.

While initial stages of deloessification occur in amazingly short time the completion of the process probably requires an indefinitely long period, possibly on the order of that needed before a soil profile attains equilibrium. Eventually loess will probably alter completely into some type of brown loam. Much of the European "*Lehm*" appears to be of such origin (Mill, 1895, p. 126).

APPLICATION OF CONCLUSIONS TO LOESS IN OTHER REGIONS

It is admittedly hazardous to apply conclusions reached in the lower Mississippi Valley to other regions. If the conclusions are sound, however, to deny their universal application is possibly unwarranted conservatism. Fact, opinion, and hypothesis are so interwoven in commonly held concepts of loess that they have become indistinguishable. The problem is presented with the hope that facts can be separated from theories and that the application of conclusions to other regions will be considered on the basis of the evidence presented.

The writer is convinced that the definition of loess used in this paper applies as well along the Rhine as in the lower Mississippi Valley. He is convinced that Lyell (1847), Hilgard (1879), and others are correct in regarding the materials as being essentially identical. His own samples, including sample 1 of Table 1, display physical identity. His field work indicates identity in origin. Deep cuts in brickyards near Aachenheim, in Alsace, reveal loess on slopes between alluvial terraces. Along the Neckar Valley, east of Heidelberg, loess also covers slopes associated with Pleistocene terraces that exhibit basal gravels, with finer materials above from which it has been

derived. At Poppelsdorf, on the outskirts of Bonn, the creep of loess over truncated edges of Pleistocene gravels is clearly exhibited, and for many miles along the foothills loess thickens downslope into valleys, where it is intensively utilized for truck crops. The same relationships occur in southeastern Netherlands, in the early Pleistocene delta region of the Rhine. At Kapel St. Rosa, 1 km. southeast of Sittard, are excellent outcrops, and many exposures reveal advanced stages of deloessification. It seems improbable that a material with such striking physical and field relationships has more than one origin. There are suggestions that it may be formed through loessification of "red clay" in China (Barbour, 1927, p. 289; Andersson, 1923, p. 107).

That there is a loess problem seems to result from two main causes: (1) the identification of many loesslike materials as loess, and (2) the insistence that the origin be eolian.

To eliminate confusion resulting from the first cause many deposits should be reexamined. The adobe of the Southwest should be dropped from consideration. Terrace deposits should be excluded. Only material such as occurs at Council Bluffs or Muscatine, Iowa, should be accepted as loess. Widespread horizontal deposits of any kind and beds containing numerous vertebrates, such as occur in the vicinity of Lincoln, Nebraska, may be suspected as not being loess. In a general way all materials not containing terrestrial snail shells should be regarded with suspicion. Acceptance of a rigorous physical definition will exclude so many loesslike deposits that the problem is reduced to relatively simple and soluble terms.

The insistence that loess be eolian has not only kept the problem alive but has also led to some amazing conclusions. Deposits of dust have been called loess in spite of almost total physical dissimilarity. The even topography of the Great Plains has been ascribed to deposition by wind, and hence surface materials have been called loess (Matthew, 1899). Elaborate hypotheses have been evoked to show that deserts to the west are sources of material entrapped by vegetation to the east (Keyes, 1912). On the other hand, loess described as occurring along the eastern sides of streams in the Central States is supposed to form "billowy ridges," more or less dune-shaped (Savage, 1915, p. 101). In the lower Mississippi Valley a topography has been created in the minds of many that does not exist in fact. It has taken the form, not of the dust dunes of Turkestan, the flats of the Great Plains, nor the billowy ridges of Illinois, but of a gigantic dunelike ridge extending from the Ohio River to Louisiana, with a high scarp facing the Mississippi and a gentle eastward slope leading to a feather edge some 10 to 15 miles away.

Barbour's (1929, p. 65) statement that loess in China "must have extended as a meniscus surface far up the mountain slopes" is an excellent topographic characterization. To explain a mantle of that sort an appeal to colluvial agencies appears more rational than any suggestions thus far provided by the group some of our European colleagues term "the aeolians."

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STRATIGRAPHY OF THE GENESEE GROUP OF NEW YORK

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ABSTRACT

The Genesee group, traced eastward from Lake Erie, changes facies in the central Finger Lakes region, New York. Black shales of western New York interfinger gradually with coarser Sherburne beds deposited on a marginal marine plain, indicating shallow water deposition for the former. Because of a temporary extension of the marginal marine plain, beds of Sherburne lithology (Starkey tongue) divide the West River shales into the Penn Yan tongue below and the Milo tongue above.

The isolithic boundary, or zone of facies change, may be extended southward to join with similar belts of change in other areas. The trend of this extension is a product primarily of the height, proximity, and trend of the highland which supplied the sediments.

In conjunction with the field study, rocks of different facies were examined in thin section for accurate determinations of rock types. Rocks are classified on the basis of size of grain and composition. Although the size of individual quartz particles ranges widely, the facies as a whole changes only one size grade, and the coarsening is due primarily to an increase in the amount rather than in the size of the grains.

A principal stratigraphic limit exists at the base of the Geneseo. There is no direct evidence that this relates to the boundary between the Middle and Upper Devonian.

INTRODUCTION

The Genesee group of beds is in the Senecan series of the New York Devonian. The simple sequence of shales in western New York becomes complicated east of Canandaigua Lake through facies changes, common to both the underlying and overlying rocks. Absence of diagnostic fossils, the thickening eastward of the mass as a whole, the thinning in that direction of units such as the West River and Geneseo, and the lack of persistent datum planes have impeded understanding of the stratigraphy. Numerous studies of restricted areas, though valuable, have not given a satisfactory conception of the regional relationships.

Reliance on paleontologic differentiation almost exclusively and in limited areas has caused confusion. Use of faunas to establish formational limits is unsatisfactory in these rocks, because the organisms tend to follow facies and cross lines of contemporaneity. This is demonstrated by the relationship between the Ithaca fauna and the Hamilton, from which the former was derived. The two faunas are separated in west-central New York by a large interval of rock, but coincide in east-central New York, where the Ithaca fauna is found in coarse-textured beds of upper Hamilton age in the Schoharie valley (Cooper and Williams, 1935, p. 826). The Hamilton fauna disappeared after Tully time in western New York but survived in the eastern part of the State. The species migrated westward following a favorable environment, but when they eventually returned to the same locality they occupied as a Hamilton fauna, they were in sediments stratigraphically far above the Hamilton. Allowing for exceptions because of irregular westward projections and tongues of coarser facies, the first abundant Ithaca fossils should be successively younger from east to west because of the time required for migration.

Convinced that facies changes are of principal importance and must be established before an understanding of the faunal shifts can be achieved, the author supplemented stratigraphic investigation by a thin-section study of the rocks in the critical interval between Canandaigua Lake and the Chenango valley. Emphasis in the field and laboratory was placed on lithologies rather than on the faunas.

FIGURE 1.—*Map of localities of New York*

FIELD WORK

Field work during the summers of 1939 and 1940 covered the area from Lake Erie to the Unadilla valley (Fig. 1). In order to trace changes and to eliminate possible error in correlating incomplete sections, the beds were measured from easily recognizable datum planes such as the top of the Tully formation. Where possible, sections were detailed along streams having complete exposures of the group.

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HISTORY OF THE NOMENCLATURE OF THE GENESEE GROUP

The beds now called the Genesee group were known and described nearly 100 years ago in reports by Vanuxem (1840; 1842) and Hall (1840). They were not referred to specifically as a group until 1933. Caster (1933, p. 202) used the term "Genesee group of formations" in listing the beds of the Ithaca region and included the Genesee (Geneseo), Genundewa, West River, Standish, and Middlesex formations. The Middlesex was subdivided into the Renwick and Six Mile shale members. Chadwick (1933, p. 95) employed the term a few months later, but enclosed it in quotation marks; in a later publication (1935, p. 310) he regarded its status as more definite. As then defined, the group consisted of only the Geneseo, Genundewa, West River, and Standish.

Vanuxem (1842, p. 168–169) applied the name Genesee to the black shale overlying the Tully limestone. Subsequently it was used to indicate the beds between the Hamilton (or Tully where present) and the Middlesex shale, the sequence being variously referred to as the Genesee beds, the Genesee shales, or simply the Genesee. In 1904, Clarke and Luther (1904, p. 25) restricted the term Genesee to the black shale below the *Styliolina* beds. In the same bulletin the term "Gorham shales" was used apparently as a synonym for the restricted Genesee formation. No definition accompanied the term Gorham, and it is of no value. Later authors used the term Genesee interchangeably for both the group and the formation. In an attempt to eliminate the growing ambiguity, Chadwick (1920, p. 118) proposed the term "Geneseo" for the restricted Genesee black shale.

The name Standish was suggested (Clarke and Luther, 1904, p. 29) to designate a transition shale and flag series between the West River and Middlesex shales in the Canandaigua Lake region. It was later extended to include the greatly thickened series of flags and shale in the Keuka-Seneca Lake area (Fox, 1932, p. 682). In this way the term was entrenched in the literature, although its validity in the area where it was originally used is questioned.

Failure to recognize the change of facies caused by a tongue of coarser rock (Starkey tongue) extending westward into the Genesee shales resulted in the erroneous designation of the beds in the Keuka Lake valley above the lower West River shales (Penn Yan tongue of this paper) as Cashaqua. This was corrected by later work in the Finger Lakes region (Fox, 1932; Bradley and Pepper, 1938, p. 8–26).

The Sherburne formation (Vanuxem, 1840, p. 381) originally referred to the series of flagstones and shales between the Geneseo and the Ithaca in the vicinity of Sherburne; they were traced as far west as Cayuga Lake. The top of the formation has been variously defined, but always on the basis of fossils whose reliability as guides in these beds is doubtful. The *Reticularia laevis* zone (Williams, 1909, p. 8) marks the top west of the Chenango valley. In the Skaneateles Lake region this zone has been named the Cornell member (Smith, 1935, p. 58, 61). The top of the formation in the type locality is still dependent on the first appearance of Ithaca fossils, as originally defined.

The initial usage of the term Portage (Hall, 1840, p. 391–392, 452–455) for the rocks between the Gardeau and Chemung in the Genesee River area was soon discarded, and the base was gradually lowered until it included the *Styliolina* lentils of the West River (Goldring, 1931, p. 369, 399). A few years later, Willard (1935, p. 1199) included the Burkett (Geneseo) and Tully of Pennsylvania in the Portage group. The United States Geological Survey recognized the Middlesex as the base of the Portage group of New York, adopting the New York Survey views of 1908 (Wilmarth, 1938, p. 1704).

PRESENT STRATIGRAPHIC NOMENCLATURE

The nomenclature of the Genesee group is simple in western New York where the clearly defined Geneseo and West River shales form the lower and upper units. Sherburne type rocks penetrate the West River of the Finger Lakes region, and the intertonguing units are named the Penn Yan and Milo tongues of the West River, and the Starkey tongue of the Sherburne (Fig. 4).

The term Genundewa is abandoned because the beds to which it refers are not considered a formation, but rather a discontinuous shell limestone facies of the West River. These lenses are referred to as the *Styliolina* or pteropod facies of the West River. Impure limestone layers overlying the Geneseo in the Seneca-Cayuga Lake region have been called Genundewa. They differ in lithology from the *Styliolina* beds to the west and are discussed as the limy facies of the Penn Yan tongue.

The name Sherburne includes the rocks between the top of the Penn Yan tongue and the base of the Middlesex remnant in the Cayuga Lake valley. Although it is apparent that part of the Sherburne is equivalent in time to the lower Middlesex, and some of the Ithaca equals the upper Middlesex in the area where the facies changes, only the black shale is referred to as Middlesex. The term Renwick shale member (Caster, 1933, p. 202) is not used, though it might well be the equivalent of the Middlesex of this paper. The designation, available only in an abstract, is not adequate for such correlation.

The term Genesee group is used to include the interval from the base of the Geneseo shale to the base of the Middlesex shale from Lake Erie to Cayuga Lake. To the east the formational divisions do not correspond to those in western New York.

For all practical purposes the Genesee group ends at Cayuga Lake. It may be correlated with the beds farther east, but cannot be extended as a definite unit, even though the lower formation continues in part to the Chenango valley.

FACIES NOMENCLATURE

An attempt to apply Caster's (1934, p. 19–36) system of facies classification to the Genesee rocks was not wholly successful. The complete succession of magnafacies as outlined by Caster for southern New York and northwestern Pennsylvania is not represented in the Genesee group. At Cayuga Lake, except for the Geneseo shale, the group is presumably within the Big Bend magnafacies, for it is lithologically like the Ithaca beds at Ithaca, New York, which Caster (1934, p. 27–28) places within that magnafacies. Westward, the rock changes to black and dark-gray thin-bedded shales which fit the description of the Cleveland magnafacies of Caster. The intervening Chagrin magnafacies, therefore, is not recognized.

The Geneseo and Middlesex black shale bands correspond to the Cleveland magnafacies; black shales of the Hamilton, notably the Marcellus, are likewise Cleveland type. They penetrate far to the east, however, into and across other magnafacies and therefore are not true parvafacies. Boundaries of magnafacies are irregular and intergradational, but such extensions as the Geneseo and Middlesex demand further clarification. If the facies are to be given names and more or less specific boundaries, the Geneseo and Middlesex black shales may be considered tongues extending eastward from the main mass of the Cleveland black shale magnafacies. The Rhinestreet black shale, higher in the Devonian rocks of western New York, may also be considered a tongue of the Cleveland magnafacies. The Starkey tongue of the Sherburne, penetrating the shales to the west, might be treated as a tongue of the Big Bend magnafacies, though its extension away from the main mass of the magnfacies is far more limited than that of the black shales. With so many variations in the ideal sequence of magnafacies and parvafacies, it is inadvisable to extend the terms into western and central New York, since they do not in themselves clarify the understanding of the relations of the beds.

GENERAL RELATIONSHIPS

Formations of the Genesee group are easily recognized by lithology in western New York. The Geneseo black shale is separated from the underlying Windom blue-gray shale member of the Moscow formation by an unconformity which increases in magnitude westward (Cooper, 1930, p. 122–123). The lithologic and faunal change at the contact is further emphasized by occasional thin lentils at the top of the Hamilton of marcasite and pyrite which contain a distinctive dwarfed fauna. In many outcrops where these lenses are apparently missing, their horizon is clearly marked by a pronounced rusty streak.

The top of the West River shale is placed at the first continuous sequence of black, fissile, brittle shale of the Middlesex formation. The Middlesex is generally distinguished from the black shales of the Genesee by the characteristic irridescent purplish-brown stain. The base of the Middlesex has been used as a plane in constructing the restored section (Fig. 2) and the stratigraphic columns (Fig. 4). No time significance is implied.

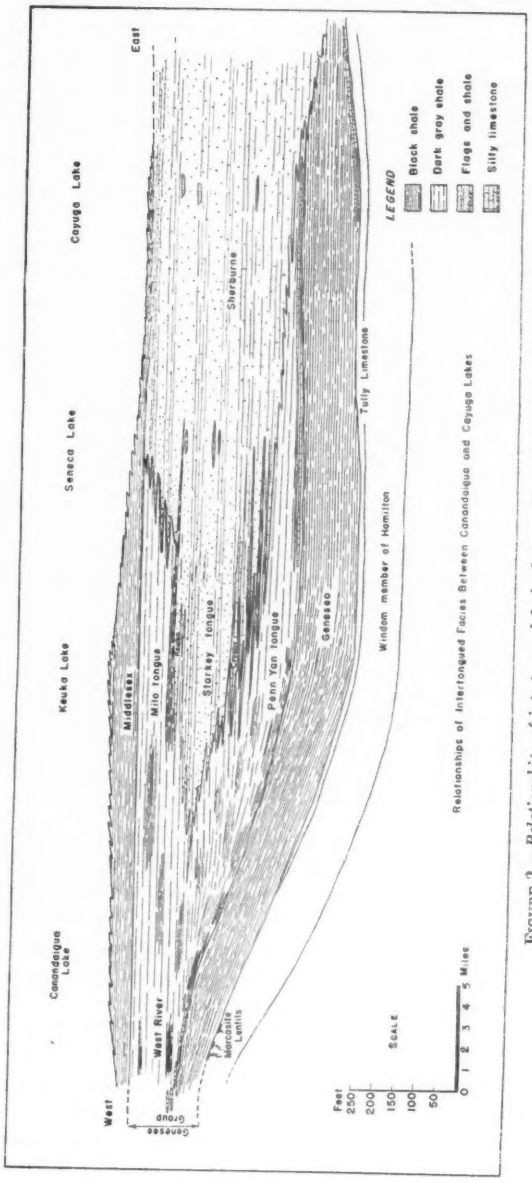


FIGURE 2.—*Relationships of interlongued facies between Canandaigua and Cayuga Lakes*

The beds of the Genesee change in the 40 miles from Canandaigua to Cayuga Lake from a dark, fissile shale to a mixture of flags, black shale, and micaceous shale with increased quartz. The eastern part of this area also marks the eastern limit of the Genesee group as a mappable unit. The Geneseo retains its black shaly character through this area, except for the eastern portion, where it contains thin argillaceous limestones at the base. The *Styliolina* facies of the West River disappears, and the base of the West River is not readily distinguished in the Keuka Lake region. An argillaceous limestone forms the base of the Penn Yan tongue of the West River in the Seneca Lake valley.

Perhaps the most significant change is noted in the sequence from the base of the West River to the bottom of the Middlesex in this area. Interbedded shales, thin cross-bedded flags, and massive silty layers penetrate westward from Cayuga Lake into the West River shales as a tongue of the Sherburne (Fig. 2).

Beginning at the east side of Canandaigua Lake, the Tully limestone rather than the Hamilton marks the base of the Geneseo and continues to underlie it throughout the extent of the black shale. The contact, like that of the Geneseo and Windom farther west, is lithologically and faunally sharp and may indicate a significant unconformity. The Middlesex continues to overlie the Genesee beds to Cayuga Lake, but at the eastern margin it is barely recognizable, appearing as only a few feet of barren, blue-black fissile shale.

The Geneseo is the only discernible unit of the Genesee group from Cayuga Lake eastward, and near New Berlin, Chenango County, even it is not recognized. Reliable indices of correlation with the beds to the west are absent. Changes of lithology east of Cayuga Lake are apparently local and do not correspond to those in western New York. Faunal differentiation is impractical because the faunas tend to reflect the lithology. Rapid changes of minor features and a general uniformity of the whole sequence characterizes the eastern equivalents of the Genesee. Unless extensive thin zones are found, there is little likelihood that formations of the group or even the group as a unit can be traced accurately east of Cayuga Lake. In this connection J. B. Reeside, Jr., (Thomas, 1934, p. 1660–1661) states:

> "It seems . . . absurd that in the case where we have no fossils and no usable lithologic differences, we should attempt a separation into hypothetical units based on tracing an invisible line from another area where the line happens to be real. That homogeneous lithologic units cover different intervals at different localities is a natural condition. It is simply a logical part of our nomenclatorial system that our names for such units have one scope at one place and another at a different place."

The statement has been used to refer to rocks of another age and area, but the situation is analagous. Although fossils are present in the Genesee equivalents, their facies nature seriously limits their usefulness.

ISOLITHIC BOUNDARY

Zones of lithofacies change, as used here, are those which separate rocks of different grain-size classification. These isolithic boundaries are somewhat indefinite and differ in position with time. The width of these zones of transition varies in successive formations and even within formations because of the complicated intertonguing of the coarse and fine layers. The geographic location of such zones, there-

fore, is necessarily general. Emphasis is placed on the belt which separates the medium siltstones to the west (essentially the black and dark-gray shales) from the coarse siltstones (flags and micaceous shales) to the east.

Because slivers of a given facies in many instances extend far beyond the main mass of that facies, the zone of facies change for each formation will be placed where one facies as a whole changes to another. There are difficulties even in such a generalized treatment. The Geneseo formation, for example, changes lithology by progressive loss of the top. The zone of Geneseo facies change must be placed arbitrarily in the Chenango valley region, where the thin remaining black shale is wholly affected by the introduction of a new facies. This is far to the east of the first difference in lithology in the upper portions of the formation.

The West River facies change is complicated by the tongue of the Sherburne (Starkey tongue) which carries the coarser lithology farther west in the middle of the formation, but allows two tongues of the West River shale to continue east (Fig. 2). The tongues of the West River do not change character at the same locality. The Penn Yan tongue has completed its facies shift in the Cayuga Lake valley, whereas the Milo tongue passes into Sherburne lithology in the Seneca Lake valley. The Starkey tongue extends westward into the Keuka Lake valley before the coarse siltstones give way to finer-grained sediments.

The facies change, if followed from the base to the top of the Genesee, forms a zig-zag line or zone. The area bounded by the eastern and western extremities of the zone, with the exception of that part of it which passes through the Geneseo, includes the Keuka-Seneca-Cayuga lakes region. The vertical isolithic boundary of the New York Genesee may be tentatively extended into Pennsylvania. This zone, when connected with the area of facies change of the fine-grained dark shales to lighter-colored and coarser siltstones in southern Columbia County, Pennsylvania, trends nearly north-south. The land supplying the sediments might have had a similar trend, though the size of grain reflects local variables such as velocity of streams and height of land mass. Current action, depth of water, slope of the submarine surface, and other marine factors similarly affect the distribution of sediments. Evidence provided by connecting only two areas of facies changes is insufficient to indicate the direction of the highland. The Genesee is not exposed between central New York and northern Pennsylvania, so this isolithic boundary must be extended farther southward before it will become useful as an indicator of the trend of the highland.

STRUCTURE AND THICKNESS

The formations of the Genesee group, like those of the underlying Middle Devonian, have a simple structure. In western New York the beds are nearly horizontal. The regional southeast dip at Lake Erie averages 35 feet per mile. In the Genesee valley the strata dip southwest at the rate of 35 to 40 feet per mile. This dip is generally constant to the Canandaigua Lake region. In the Keuka Lake valley the dip becomes more westerly but shifts to the south again in the Seneca Lake valley where it is between 20 and 30 feet per mile. The southward direction is continued into the Cayuga Lake valley and increases to 40 feet per mile. At Moravia it swings

back to the southwest and maintains this direction into the Chenango valley, gradually steepening to 70 feet per mile.

Affected by innumerable small folds, the dip in some areas changes many times within very short distances. Some larger flexures and faults have been mapped, particularly in the area between Canandaigua and Cayuga lakes. A dome with a closure of from 50 to 80 feet has been reported in the Barrington and Milo townships on the east side of Keuka Lake (Fox, 1932, p. 686–688). Bradley and Pepper (1938, Pl. 2) report a faintly defined structural plateau in the Seneca-Keuka lakes region, outlined by south-dipping rocks a short distance south of Seneca Lake and by northwest- and west-dipping beds along the Keuka Lake valley.

The existence of the Clarendon-Linden fault (Chadwick, 1920, p. 117–120; 1932, p. 143) or monocline described as trending north-south just east of Batavia and presumably offsetting the beds approximately 100 feet is questioned (Hamilton, 1937, p. 1586). In the Keuka Lake outlet a fault having a throw estimated to exceed 60 feet is well shown by the repetition of the Tully limestone which underlies the Geneseo shale. Several small faults cut the strata along the shore of Seneca Lake; the largest is the Pioneer Camp fault on the west side, where the displacement is more than 40 feet (Fox, 1932, p. 688). Bedding faults are not uncommon but are recognized with difficulty because they produce little apparent disturbance. Other smaller faults and folds continue to the east, and the whole series has been assigned to the Appalachian system (Kindle, 1904, p. 281–289).

Well-defined jointing in many cases controls the drainage for short distances. Two major joint systems strike N. 10° W. and N. 85° W. There are, of course, variations in these directions of the master sets over the distance from Lake Erie to the Chenango valley. Numerous other variously oriented joints are present. Some are so closely spaced in many places in the shales that the rock appears to have been diced. Curved joints are relatively rare.

Ripple marks are common and range in wave length from a fraction of an inch to 1.5 feet. These are apparently haphazardly oriented and are confined to the coarser layers. Both current and oscillatory ripple marks are present. Small-scale cross-bedding in the coarse layers throughout the sequence is commonly associated with ripple marks. Weathering proceeds along cross-bedded planes in some of the flaggy layers and tends to produce uneven surfaces which may be mistaken for ripple marks. The shallow water origin of the cross-bedding and ripple marks is emphasized by the presence in adjacent rocks of cross-bedding on a larger scale, formed presumably by swifter currents. Other structures include intraformational contortion, channel filling, and accumulations of marcasite pellets (Pl. 2, fig. 3). The large-scale cross-bedding and its associated features are well displayed by the relatively coarse-grained (coarse quartz siltstone) Sherburne beds in the Trumansburg section on the west side of Cayuga Lake.

Grooves and striae in parallel arrangement and haphazard distribution indicate either ice-flow markings (Clarke, 1918, p. 199–238) or drag of debris by current and wave action. These again suggest very shallow water. The occurrence of the coarse, branching, tubular "*Fucoides graphica*", though more prominent in higher beds, presents a problem of origin which has not yet been solved.

Flattened, shiny, thin disc-shaped clay galls averaging one half to three quarters of an inch in diameter are prevalent in local areas of the Starkey and Sherburne. They are confined to the coarser layers, and many are associated with oscillatory ripple marks. Their occurrence in the siltstones strongly suggests very shallow water or even subaerial sedimentation.

Marcasite nodules throughout the shales and siltstones vary in diameter from a fraction of a millimeter to half an inch. A local accumulation of spherical marcasite pellets occurs with abundant badly macerated plant remains in the Sherburne of the Trumansburg Creek section. These resemble oölites but lack characteristic oölitic structure. They possess a definite core, however, which is shattered in many and separated from the main mass in others (Pl. 2, fig. 3). The margins of some of the spheres are discolored as if weathered. Irregular masses of marcasite appear in the same rock with the pellets. The presence of spores in associated beds and the abundance of other plant remains, coupled with the comparative uniformity of size of the pellets, suggests that these spheres represent replaced plant spores. The irregular masses are merely accumulated iron sulphide that filled interstices between the pellets and quartz grains.

Concretions are the most abundant sedimentary structures. These range in size from a fraction of an inch to several feet and in shape from spheroids to elongate cylindrical masses. They are accumulations of argillaceous limy material, and many contain small crystals of pyrite, barite, galena, sphalerite, calcite, sulphur, gypsum, or quartz. They rarely contain fossils. Concretions occur singly and in rows throughout the group but are particularly abundant in the West River shales where the elongate type reach an observed maximum length of 11 feet. They are commonly cut by joints, indicating that they antedate the stresses which produced these joints. Some of the shales surrounding the concretions are squeezed and warped. Many of the concretions were undoubtedly contemporaneous or penecontemporaneous in origin, since they were formed before the Applachian revolution to which the joints are assigned. They show no signs of further growth after being cut by the joints, and it is likely that they did not enlarge after induration. It is sometimes confusing, however, to assign relative ages to the concretions. Some which are cut by the joints are also characterized by bedding planes, continuous with the adjoining rocks. These are probably epigenetic, and the fact that they are cut by joints may indicate movement, after growth of the concretions, along previously established joint planes. Those concretions which occur in rows may be useful in local correlation.

ROCK TYPES

GENERAL STATEMENT

Thin-section study of sedimentary rocks is a valuable supplement to field investigation. Microscopic examination of 75 thin sections representing rocks of several facies and of wide areal extent was undertaken to establish a more accurate quantitative basis for the classification of the Genesee rock types. Studies were made of the sizes of quartz grains in all the rock types and of relative percentages of the major rock constituents.

Specific descriptive terms are used to clearly differentiate the facies. The grade-size classification employed in the thin-section work is based on a convenient decimal scale, which compares closely in most respects with the Wentworth scale (Fig. 3).

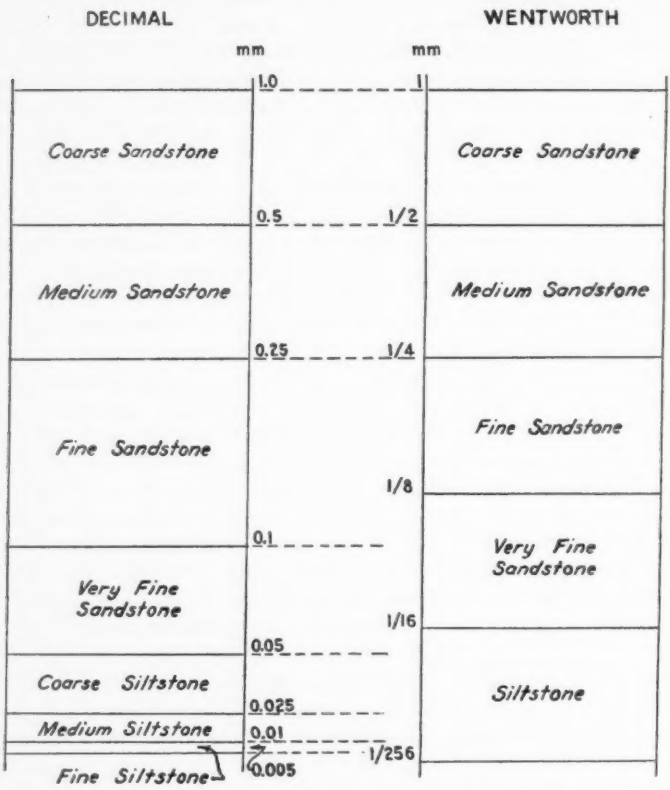


FIGURE 3.—*Diagrammatic correlation between the decimal scale and the Wentworth scale*

The size grade of a given rock is based on the sizes of the quartz grains. Quartz is used as a standard to avoid discrepancies due to flocculation or alteration which might affect the sizes of other minerals. It must be emphasized that the factor of percentage of quartz is equally as important as size, because a rock of the medium siltstone grade with only 10 per cent of quartz cannot be considered the same facies as one of equal size classification containing 30 per cent quartz. Thus, each size classification must be qualified by an adjective denoting composition.

Rocks of the Genesee group, on the basis of the size of the mode, grade from black, highly argillaceous fine siltstone to very fine quartz sandstone (Pls. 1, 2). Individual grains reach the fine sandstone grade, but the proportion of these larger

grains to those of the mode is extremely small. Rocks of all facies contain an abundance of the very fine size grades.

Thin sections give a clearer conception of the intergradation of rock types. The tendency for a general increase in quantity and size of the quartz grains eastward is interrupted by considerable local variation due to the interfingering of the facies. In assigning a given facies to any classification, the dominant type must be used, for any given facies has within it rock types characteristic of other facies. In general the change of facies of the Genesee from west to east as far as the Sherburne valley involves a shift of only one size grade—from medium to coarse siltstone. No widespread sharp break is present between the facies of the Genesee group either laterally or vertically, but rather a gradation or interfingering of types occurs, with the eventual predominance of one type. Some sharp changes are noted, such as that exhibited by the lower Starkey (Crosby lentil) of the Keuka Lake area, but these are few and only of local importance. Increase in percentage of quartz fragments generally seems to precede the shift to larger sizes in the intertonguing beds. The coarsening is due primarily to greater quantities of quartz grains rather than a spectacular increase in the size of the grains themselves, since the actual size difference of the mode between the finest and coarsest rock types of the Genesee group is only 0.065 mm.

METHOD OF PROCEDURE

Alling's (1941, p. 28–31) method was used to measure and count grains in thin section. The apparatus consists of a metal diaphragm and a camera lucida attached to a petrographic microscope. The diaphragm is mounted over a hole in the camera lucida drawing board through which a colored light shows. A transparent scale is placed just below the diaphragm and over the hole. These scales correspond to each combination of objective and ocular of the microscope and are marked off in concentric circles reading in fractions of a millimeter.

The image of the scale is superimposed on the grain by means of the camera lucida, and the diaphragm is adjusted until the area of the superimposed circle outside the grain is equal to the area of the grain not covered by the colored circle. When this is brought about, the diameter of the circle, representing the nominal sectional diameter of the grain, is read directly from the scale. The method gives the same results as the use of the planimeter for measuring grains, is far more rapid, and is accurate.

A count of 200 grains per slide is a reliable sample. It gives a minimum statistical error and is not too high to seriously reduce the speed of the total operation. A comparison of results from counts of 50, 100, 150, 200, 250, 300, 350, and 400 grains on a number of slides shows that the mode shifts into different sizes with each count up to 150 grains. There were only occasional differences in the sizes of the mode between counts of 150 and 200 grains. The size of the mode resulting from a count of 200 grains and that resulting from higher counts changed only in one instance. This figure compares favorably with that suggested by Krumbein (1935, p. 493), who indicates that a count of about 250 grains is satisfactory when the planimeter method is used. Figures derived from the thin-section study must be adjusted before

assigning a rock to a given size classification, since a random section through grains results in a reduction of the actual grain size. The extent of this reduction varies with the shape of the grains. The average reduction of angular grains, such as those of the Genesee group, is approximately 20 per cent (Alling, personal communication, 1941).

The relative percentages of the constituents of the rock were determined by Wentworth stage analyses.

STRATIGRAPHY

GENESEO SHALE

General considerations.—Most widespread unit of the Genesee group is the Geneseo shale, the basal formation. It extends from Lake Erie to Cayuga Lake, beyond which its black shale facies is recognized as far as the Chenango valley (Fig. 4). East of Cayuga Lake the upper portion of the Geneseo grades laterally into the coarser facies characterizing the higher beds; the black shale facies thins and finally disappears a short distance east of the Chenango valley.

Chadwick (1920, p. 118) named the Geneseo shale and suggested as the type locality the section at Fall Brook, $1\frac{3}{4}$ miles south of the town of Geneseo. The entire formation is revealed here, but careful examination of the whole sequence is difficult because it lies high in the walls of the ravine and in the face of a sheer falls. Other near-by outcrops, however, can be studied. The formation is 84 feet thick here. The Geneseo thickens from a 2-inch sliver at Lake Erie to 125 feet at Cayuga Lake, but thins eastward from this meridian to 40 feet in the Chenango valley; it disappears by intergradation with coarser rock to the east. The unit is characteristically black, brittle, massive, finely laminated shale, becoming fissile on weathering. It contains interbedded gray shales, thin limestone lentils, irregularly distributed concretions, and marcasite nodules. The lower 10 to 20 feet of the formation is uniformly densely black and makes a sharp contact with the underlying beds. Distribution of the lighter-colored shale, limestone lentils, and the concretionary zones varies. The lighter-colored portions of the Geneseo closely resemble those of the overlying West River, making the two formations nearly indistinguishable in the stream flowing into the Keuka Lake outlet at Seneca Mills, where the *Styliolina* facies of the West River is missing and the limy facies that characterizes the lower part of the West River to the east is not clearly defined. In such a locality the top of the Geneseo is arbitrarily established. The main mass of the overlying West River is recognized by the fairly abundant pelecypod fauna, but the contact between the two formations is gradational.

The Geneseo is predominantly a medium shaly, argillaceous siltstone with a small amount of fine siltstone (Pl. 1, fig. 1, 2). Quartz grains are angular and have ragged edges. The average of the modes is 0.018 mm. in diameter. Some grains are less than 0.005 mm., and a few reach 0.10 mm. in diameter. Hard (1931, p. 172) states that the average diameter of the quartz grains of the Geneseo of central New York is 0.03 mm., but his figure is based on the average of all the grain sizes rather than the average of the modes (Hard, personal communication, 1941).

Quartz particles are usually scattered in a groundmass of minutely crystalline,

flaky material classified as clay, which is strongly colored by finely disseminated carbonaceous matter. The abundance of this organic material obscured the slides in many instances, so it was difficult to find quartz grains with outlines clear enough for accurate measurement. In addition, larger fragments of plant remains and occasional flattened plant spores, oriented parallel to the bedding, were present. Cryptocrystalline marcasite is commonly disseminated throughout the rock.

The base of the Geneseo in the Cayuga Lake-Moravia-Skaneateles Lake region is composed of very fine-grained somewhat argillaceous limestone (Fig. 4). These layers contain over 80 per cent calcite and approximately 2 per cent quartz. A sufficient number of grains could not be measured to establish a reliable mode, but no grains exceeded 0.03 mm. in diameter. The mode size is fine siltstone. Marcasite and carbonaceous material are present. Four of these layers, each averaging about a foot thick, are separated by thin shale bands in the Taughannock Creek section of Cayuga Lake, the whole unit occupying 6.8 feet. In the Moravia area to the east, the 13 feet overlying the Tully limestone is composed of black, argillaceous limestone with few shaly partings. Three thin limy bands occur at the base of the formation in the Skaneateles Lake valley but disappear again to the east.

Progressive eastward displacement by coarser siltstones involves the loss of the top of the black shale east of Cayuga Lake, though in certain areas the contact appears clearly defined. The change in color seems to precede the shift to a coarser facies, however, so that some of the lighter-colored beds at the top of the Geneseo are still in the medium siltstone class. The percentage of quartz grains increases noticeably in these lighter-colored beds. They contain up to 23 per cent quartz and are classified as medium argillaceous siltstones. Interbedded layers of medium quartz siltstone contain as much as 43 per cent quartz. The change in facies, then, begins with the introduction of thin bands of rock with more quartz but no increase in the mode size of the grains. These thin layers increase in number and thickness eastward, and the size of the mode also increases until the change is completed and the rock has graded into a brownish, coarse siltstone. Mica is present in increasing amount toward the source of the sediment.

The remaining black shales change gradually from Tully eastward. Though easily recognizable, the shale becomes interspersed with coarser and lighter-colored layers, resulting in an interfingering with the main mass of coarser beds above. In its easternmost outcrops, the Geneseo is somewhat lighter and definitely silty, though still thin-bedded, and has some interbedded black shales. The 40 feet in the Sherburne valley is mainly medium siltstone with interbedded coarse siltstone.

Farther east, in the vicinity of New Lisbon, the rocks which occupy the same stratigraphic position as the black shale of the Geneseo are brownish gray and are predominantly micaceous coarse siltstones containing grains as large as 0.095 mm., some of which appear to be slightly rounded. The quartz content in these rocks is 50 per cent or more. Interbedded with these strata are many medium quartz siltstones strongly resembling the lighter-colored layers in the Geneseo shales to the west, with their relatively high percentage of quartz and small size of mode.

The Geneseo is quite fossiliferous, though the distribution of the fauna tends to be spotty and the variety limited. The flora is abundant in places, but usually too

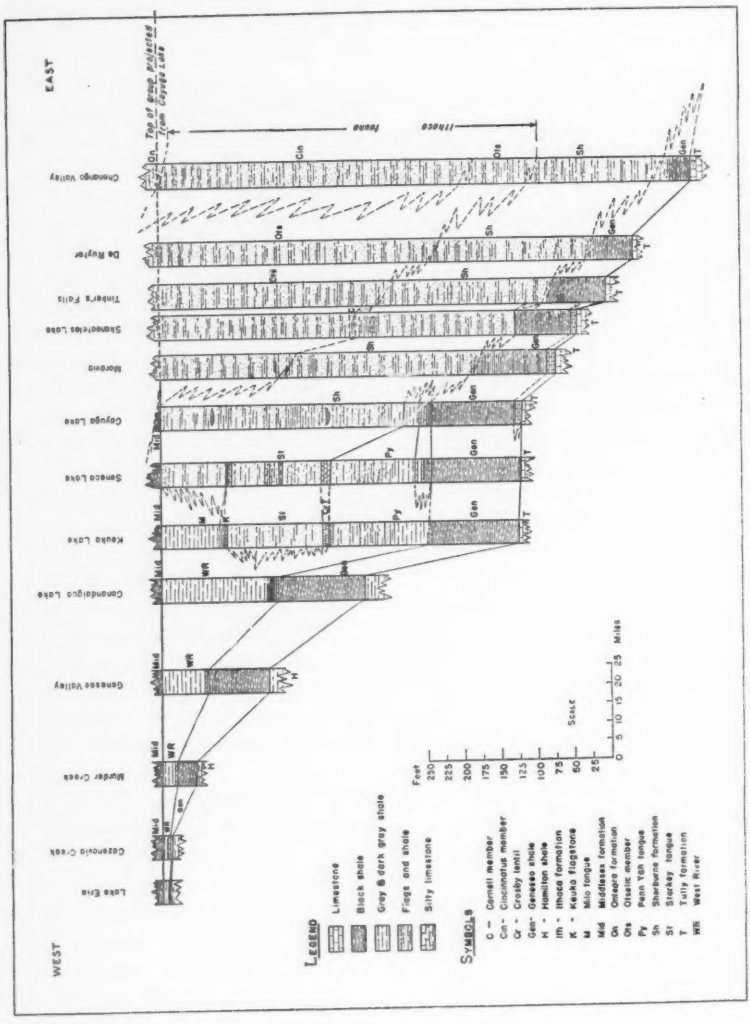


FIGURE 4.—*Columnar sections between Lake Erie and Chenango valley, New York*

poorly preserved for proper identification. The fauna includes a few Hamilton forms which continued to flourish despite the influx of suffocating muds of the Geneseo. Some of the forms, particularly *Leiorhynchus*[1], are predominantly black shale fossils and can be found in black shales as far down as the Marcellus at the base of the Hamilton. Common fossils of the Geneseo are: *Leiorhynchus quadricostatus*, *Orbiculoidea lodiensis*, *Lingula spatulata*, *Chonetes lepidus*, *Styliolina fissurella*, *Pterochaenia fragilis*, *Bactrites aciculum*, and *Probeloceras lutheri*. Found occasionally are *Buchiola retrostriata* and *Pleurotomaria* (*Euryzone*) *rugulata*. Careful search revealed only a few poorly preserved specimens of *Schizobolus truncatus*, only one of which could be identified with reasonable assurance. One specimen of the large pelecypod *Panenka robusta* was found in the Genesee valley. The fauna is nondiagnostic, contains elements of both Hamilton and Naples faunas, and may be considered transitional between the two.

No persistent faunal zones have been established in this formation, though a 15-foot interval at the top, in the Seneca-Cayuga Lake area, contains abundant *Orbiculoidea* and *Leiorhynchus*. This zone was not recognized, however, in the Keuka Lake area, and in the Canandaigua Lake region these fossils are nearer the base of the formation than the top.

Upper and lower contacts of the Geneseo.—The line of division between the Middle and Upper Devonian cannot be established in New York. Evidence is insufficient to warrant attaching regional importance to disconformities in the Genesee and associated strata, although there is a principal stratigraphic boundary at the base of the Geneseo. The subsequent discussion details facts and interpretations pertaining to the relationship of the Geneseo to the beds above and below.

The Geneseo-Tully contact assumes increased importance if the Tully is classified as Middle rather than Upper Devonian. If the Geneseo and Tully are both Middle Devonian, the upper contact of the Geneseo gains relative importance.

Grabau (1917, p. 951) states that the Tully grades into the Geneseo and that the Geneseo replaces the Tully by progressive overlap. The Tully in some localities has shaly upper beds and may be considered to grade into the overlying Geneseo shales. Ordinarily, a color change is noticeable, and character of bedding differs, however, so that the contact is recognizable. Where the Geneseo has limestone at the base, the contact might be called intergradational, but the difference in character of the limestone permits the two formations to be separated readily. Lithologic distinction is possible, therefore, between the Geneseo and Tully except where both change to a coarser facies in the eastern part of the State.

In Pennsylvania the Tully and Geneseo have been treated as a unit. The Burkett shale, presumably the Pennsylvania correlative of the Geneseo, has been placed with the Tully in the Portage group as a member of the Rush formation (Willard, 1935, p. 1199). The Tully in Pennsylvania apparently grades into the overlying shales by interfingering of the shales and limestone. The Burkett is "divided from the overlying Harrell shale by a sharp lithologic change (practically a break) accom-

[1] Although characteristically a black shale habitué, *Leiorhynchus* does not occur in the Middlesex black shale higher in the sequence. Conversely, it does occur in considerable abundance in the silty, light-colored beds of the Sherburne and "Ithaca". It is present also in the limy facies at the base of the West River in the Seneca Lake area.

panied by the appearance of plentiful Naples fauna" (Willard, 1935, p. 1209). A somewhat similar relationship is noted between the Geneseo and the *Styliolina* facies of the West River, which occupy the same respective positions; but the faunal differences are controlled by facies, and where the *Styliolina* beds are absent, the change is gradational. A widespread submarine disconformity might be indicated by the *Styliolina* beds, on the assumption that the deposition of great multitudes of *Styliolina* shells involved a very long time. The significance of these lentils in defining a continuous break in western New York is doubtful, however, because of the highly irregular and only local occurrence of the pteropod lenses.

The contact between the Geneseo and the overlying Sherburne in the Skaneateles Lake area is sharp. The Geneseo shale is succeeded there by a heavy siltstone bed which Smith (1935, p. 54) recognizes in many places on the Skaneateles quadrangle. The writer found occasional fragments of the fine-grained black Geneseo within the base of the Sherburne in this area. This is a minor disconformity in a shallow-water deposit and has no regional significance.

The inarticulate brachiopod *Schizobolus truncatus*, considered a guide fossil for the Geneseo, has been reported from the upper Hamilton shales (Cooper, personal communication, 1941) but not from the strata above the Geneseo. This may indicate that the Geneseo is Middle Devonian. The value of *Schizobolus* as a guide for the Geneseo may be questioned, however, because its rare occurrence in this formation may indicate that its apparent absence in younger beds merely reflects less thorough search. Extreme caution must be exercised in the use of long-range fossils of this character in indicating time relationships.

Placing a break at the top of the Geneseo is supported by the occurrence in the *Styliolina* layers of an abundance of fossils, many of which are new, and most of which are found in the higher beds. The pteropod layers are limestone, however, and it is not surprising that organisms are more abundant and different from those in the black shales below. Facies control is evidenced by the fact that fossils are more varied and abundant principally in the limestone layers, and the separating shales carry a typical West River biota; also, as indicated previously, the new forms were not found in the shales which occupy the level of the *Styliolina* layers where these are missing in the Keuka Lake region. The introduction of a different cephalopod fauna in the *Styliolina* facies distinguishes it faunally from the Geneseo. Many of the species of *Manticoceras*, characteristic of the Naples fauna, begin in the pteropod limestone. Related forms, however, are found in the beds below the *Styliolina* lentils. Miller (1938, p. 10) lists the doubtful form *Manticoceras* (?) *amplexum* from the Tully limestone. Cooper and Williams (1935, p. 859) list *Tornoceras uniangulare* from the Tully. The latter form is found also in the pteropod facies. *Probeloceras lutheri* is rather common in the Geneseo. Identification is difficult, however, because the fossils are commonly flattened, distorted, and the all-important sutures have been lost. *Probeloceras* and *Manticoceras* resemble each other, moreover, in many stages of development. Miller (1938, p. 72) states that "*Manticoceras* differs from . . . *Probeloceras* in that each of its complete sutures forms a total of six, rather than four, primary lobes". The badly crushed forms in the Geneseo, lumped into *Probeloceras* for want of proper features to aid accurate identification, may represent forms

of *Manticoceras*, and the appearance of the forms first in the *Styliolina* facies is a matter of finer preservation in the calcareous rocks.

Miller lists eight species of *Mantioceras* from the *Styliolina* beds; one additional form is of doubtful occurrence there. Of these, four are restricted to the pteropod facies, and only a single specimen of one of the remaining species has been reported in the higher beds. The characteristically rapid changes of this type of organism are well known, but the extremely limited range of many of the species of *Manticoceras* hints at facies control of these forms.

The faunal alteration between the Geneseo and overlying strata is of less importance than that between the Geneseo and Tully in west-central New York. Farther east where the Tully and Geneseo equivalents have essentially the same facies, they carry similar faunas, but these are found in overlying rocks as well.

The sharp change from Tully limestone to Geneseo shale is responsible to some degree for their dissimilar fossils. Where the lower Geneseo is limestone, however, the faunal change is still maintained. Also, where the top of the Tully is shale, Tully fauna continues into this changed facies, ending at the base of the Geneseo. At Kashong Creek on the west side of Seneca Lake the top 6 inches of the Tully consists of calcareous shale with abundant corals; 3 feet of calcareous shale carrying *Phacops* (Cooper and Williams, p. 795, 805) forms the top of the Tully at West Brook, in the Chenango valley (eastern extremity of the limy Tully). These fossils, of Middle Devonian aspect, are missing from the Geneseo shale. Therefore, the principal faunal changes are not controlled by a change of facies. The importance of the effect of facies changes alone on the Hamilton-type fossils is questionable. Closely related forms are abundant in the marcasite lentils at the top of the Hamilton in western New York. The forms are dwarfed, but they nevertheless continued to flourish in an exceedingly unfavorable environment. Species with close Hamilton affinities lived through three different environments, represented by Hamilton shale and marcasite and Tully limestone. The sudden termination of all but a few of these organisms at the base of the Geneseo, then, appears to indicate something of greater significance than a change of facies. It is suggested that an important disconformity exists here. Proof of whether or not it represents the line of demarcation between the Middle and Upper Devonian depends on correlation of the Tully and Geneseo with the Givetian and Frasnian stages of Europe.

Further evidence of such a disconformity is shown at Taughannock Creek. The top of the Tully is irregular with some scoured channels half an inch deep. These are distinguished from the narrow stylolitic joint depressions which are present. The top surface of the Tully bevels the small corals *Lopholasma*. These features are not of recent origin, inasmuch as they are absent in the lower beds of the Tully whose surfaces are similarly exposed. A few limestone pebbles at the contact further emphasize the disconformity.

The Geneseo is regarded as part of the Upper Devonian as presently defined.[2]

[2] The Genesee and all the New York Upper Devonian are part of a flysch deposit which begins after the close of the Onondagan. Division of the Devonian into two rather than three major units thus may be indicated. On this basis, strata to and including the Onondaga comprise the Lower Devonian; all post-Onondaga beds are Upper Devonian; the term Middle Devonian would be eliminated. Such a classification does not affect the validity of the series terms such as Erian, Senecan, and others.

WEST RIVER FORMATION

General statement.—The West River formation, as first described (Clarke and Luther, 1904, p. 28), consists of interbedded blue-black, dark-gray, and black fissile shale with abundant spherical concretions. It is proposed here to include the *Styliolina* limestone facies (Genundewa limestone) in the base, so that the lower boundary of the West River will be the first or lowest *Styliolina* layer. The upper boundary of the formation is marked by the black Middlesex shale. The West River continues as a clearly recognizable unit from its type area at Canandaigua Lake to Lake Erie (Fig. 4).

Styliolina *facies.*—The pteropod limestone lentils at the base of the West River are the most distinctive beds of the Genesee group. The *Styliolina* facies consists of thin lenticular deposits of shell limestone and interbedded dark shale, outcropping from Lake Erie to Canandaigua Lake. Shells of *Styliolina fissurella* are concentrated in enormous numbers in the limestone layers and are distributed through the intervening shales. Elliptical concretionary masses of *Styliolina* occur in the limestone bands, and some are in the shale between. Many are set in thin, platy, carbonaceous shale bands which are compressed below the nodules and curve around them much as do the shales enclosing the larger argillaceous concretions of the Geneseo and West River. These nodules are probably penecontemporaneous and resulted from concentration of very thin layers of saturated limy mud containing an abundance of *Styliolina* shells. Loss of water shortly after deposition caused the accretion of the shells into spheres forcing the mud enclosing these balls to curve around them.

The shells lie at all angles to the bedding. Relatively few are broken, and the collapse of the cone-shaped shells was probably due to compaction. Quartz grains are rare, and all fit into the medium siltstone size. Calcite averages between 80 and 90 per cent of the rock, and most of it is recrystallized. It forms the shells of the organisms and fills the interstices. The *Styliolina* facies contains in places as much as 10 per cent of carbonaceous material, much of it the remains of the soft parts of the pteropods. In addition there are occasional plant spores and fragments of plant remains. Bituminous material has pervaded the rock and given some parts a strong odor and a brownish tinge when freshly broken. Marcasite replaces some of the carbonaceous matter in the *Styliolina* shells and fragments of plant remains. Pyrite is also present in minor quantities. The pteropod facies is the only rock of the series in which consistent occurrence of cubic pyrite crystals was noted, though they occur in some concretions in the other strata.

A few glauconite grains are present in the limestone, but the total amount of the mineral for each slide is less than 0.5 per cent. Glauconite is thought to originate in relatively shallow water as the alteration product of biotite (Galliher, 1935, p. 1351–1366). Its association with areas of little agitation, black-mud bottoms, and anaerobic conditions fits with the evidence of the *Styliolina* deposition. The small amount of glauconite is due probably to the original scarcity of biotite.

The *Styliolina* beds are considered a facies of the West River rather than a separate formation. Individual layers are lenticular and range in thickness from a fraction of an inch to 1 foot; the *Styliolina* facies as a whole ranges from a few inches

to 46 feet. The variation, however, is not related to direction, as it tends to be in the overlying and underlying beds. Spectacular changes in thickness of this facies are well shown, particularly in the streams on the east side of the Honeoye valley, 1½ and 2 miles north of Honeoye, where the interval between the bottom and top limestone is reduced from 46 feet to 7.5 feet in less than half a mile. This involves the loss of some of the layers, but in some instances one section with few layers is thicker than another with a greater number of *Styliolina* layers. Exposures at Beards Creek, Little Beards Creek, and the stream just south of Teeds Corners in the Genesee valley (Caledonia quadrangle) illustrate this feature. With such variations in thickness of the whole unit, the thickness of the individual bands, and in the number of limestone bands, correlation of separate layers is inadvisable. Chance occurrence of this type of limestone is demonstrated by the discovery of a thin isolated lens in the Starkey tongue of the Sherburne at Willow Grove glen on the east side of Keuka Lake; it differs from the lower pteropod layers only in that the quartz grains are larger (Pl. 1, fig. 4), and the shells are oriented parallel to the bedding.

Shale layers between the *Styliolina* lentils are lithologically and faunally allied to the West River. They consist of black and dark-gray shales and contain such common West River forms as *Buchiola retrostriata*, *Pterochaenia fragilis*, *Paracardium doris*, and *Styliolina fissurella*.

Clarke and Luther (1904, p. 59) recognize 43 forms from the *Styliolina* layers of the Canandaigua Lake area:

Dinicthys newberryi
Echinocaris (?) *longicauda*
Spathiocaris emersoni
Cardioceras
Manticoceras styliophilum
M. contractum
M. fasciculatum
M. nodifer
M. (*Gephyroceras*) *genundewa*
Tornoceras uniangulare var. *compressum*
Orthoceras atreus
Styliolina fissurella
Tentaculites gracilistriatus
Gyroma (*Pleurotomaria*) *genundewa*
Bucanopsis (*Bellerophon*) *koeneni*
B. (*Bellerophon*) *denckmanni*
Buchiola livoniae
B. scabrosa
Paracardium doris
P. delicatulum
Lingula spatulata
Lingulipora williamsiana
Phragmostoma natator
P. incisum
Loxonema noe
Macrochilina pygmaea
M. seneca
Diaphorostoma pugnus
Protocalyptraea styliophila
Chaenocardiola (*Lunulicardium*) *hemicardioides*
Lunulicardium encrinitum
Pterochaenia sinuosa
Honeoyea styliophila
H. simplex
Ontaria suborbicularis
Buchiola retrostriata
Aulopora annectens
Cordaeoxylon clarkei
Cladoxylon mirabile
Cyclostigma affine
Lepidodendron gaspianum
L. primaevum

Many of these reappear in the Cashaqua. Their absence in the black and dark-gray shales of the West River and Middlesex and in the Geneseo below points to facies control of the fauna. It is true, of course, that the *Styliolina* facies is limy, and the Cashaqua is shaly, but the Cashaqua is a gray and olivaceous series of shales and appears to have provided a more suitable environment for the existence and development of these forms than did the dark muds of the Geneseo and West River.

Basal limestone facies of West River of Seneca-Cayuga Lakes area.—Silty limestones and interbedded shales, occupying approximately the same stratigraphic position

as the *Styliolina* facies, occur in the Seneca-Cayuga Lake area (Figs. 2, 4). The unit is 23 feet thick at Lodi Glen, on the east side of Seneca Lake, and consists of dark blue-gray shales and limestones grading into one another both vertically and horizontally. Calcareous concretions ranging from 1 inch to 1 foot appear in the shale. The contact with the West River above is in places difficult to locate, but the contact with the underlying Geneseo is sharp. These layers have been referred to as the Genundewa limestone. The limestones and shales contain abundant West River fossils, but the predominance of *Styliolina*, which distinguishes the "Genundewa", no longer is maintained.[3]

The limestones and shales are more silty in the Cayuga Lake valley. They are 16 feet thick in Trumansburg Creek, and both contacts are distinct. The layers grade into each other essentially as in the Seneca Lake region. Some of the limestones tend to occur as concretionary bands. The number and variety of fossils is materially reduced in comparison with the beds in the Seneca Lake valley. These beds lose their identity east of Cayuga Lake.

The series contains an abundance of fossils in the Seneca Lake region, though it has considerably less variety than the *Styliolina* layers. The assemblage is more normal to Geneseo and particularly West River than is the *Styliolina* fauna. The list furnished by Clarke and Luther (1909, p. 27) includes:

Manticoceras patersoni
Bactrites sp.
Gomphoceras cf. *manes*
Paleotrochus praecursor
Loxonema var.
Styliolina fissurella
Buchiola retrostriata
Palaeoneilo muta
Pterochaenia fragilis
Pleurotomaria capillaria
Ambocoelia umbonata
sp. cf. *subumbonata*
Chonetes scitulus
Leiorhynchus mesacostalis
L. globuliformis
Orthothetes sp.
Orbiculoidea lodiensis
O., small form
Lingula spatulata
Cladochonus, abundant in the concretions

West River shale.—The main mass of the West River, exclusive of the pteropod facies, maintains a relatively uniform lithology in western New York. It resembles the underlying Geneseo formation, especially in the lower part, although it tends to be more limy throughout. In the region from Murder Creek westward, the shale is black and fissile and as such is the counterpart of the Geneseo shale below.

The shale thickens from 8.5 feet at Lake Erie to a maximum of 165 feet at Canandaigua Lake; in the intervening area the thicknesses are: C zenovia Creek, 10.8 feet; Murder Creek, 27.5 feet; Genesee valley, 60–70 feet (Fig. 4).

The West River in western New York, like the Geneseo, is predominantly medium argillaceous siltstone. The mode of the quartz grains averages 0.013 mm. and places the shales close to the lower limit of the medium siltstone grade. The shales are characterized by a high percentage of clay and usually less than 10 per cent and not more than 14 per cent of quartz. Calcite ranges from less than 10 per cent to 31 per cent. Marcasite and carbonaceous matter are present in much the same proportion as in the Geneseo.

[3] Bradley and Pepper (1938, p. 10) state that the limestone at Lodi Glen contains an abundance of *Styliolina* and is therefore the typical Genundewa limestone. The author failed to find *Styliolina* in sufficient quantity to regard these layers as typical pteropod facies.

Numerous thin, shaly, calcareous argillites and large numbers of limy concretions are interbedded with the shale. These limy aggregates vary from small spheres only 1 inch in diameter to great, elongate, irregular masses up to 11 feet long. They form layers which are persistent for only short distances and occur scattered singly throughout the shales. Some contain fossils which are "more in accord with the singular fauna of the Genundewa limestone than are those of the shales" (Clarke and Luther, 1904, p. 28), and support the idea of facies control of the *Styliolina* fauna. Septaria are not common, but some were found in the Genesee River and Conesus Lake valleys.

The thin-shelled pelecypod, *Pterochaenia fragilis*, is the most abundant fossil in the West River. Nondiagnostic, it is found also in the Geneseo and extends down to the Oatka Creek black shale at the base of the Hamilton. Fossils common in the West River shales, in addition to *Pterochaenia fragilis*, include: *Chonetes lepidus*, *Pleurotomaria rugulata*, *Buchiola retrostriata*, *Paleoneilo constricta*, *Paracardium doris*, *Ontaria suborbicularis*, *Bactrites aciculum*, *Manticoceras sinuosum*, *Probeloceras lutheri*, *Styliolina fissurella*, crinoid stems, and plant remains. *Melocrinus clarkei* occurs in a thin layer near the base of the shales in the Canandaigua Lake area (Clarke and Luther, 1904, p. 59–60). Some Geneseo forms such as *Lingula spatulata*, *Orbiculoidea lodiensis*, and *Panenka* are found occasionally in the West River. The fauna is made up of individuals which, with others not found in the West River, constitute the Naples fauna.

The top of the West River at Wilder Run, near Bristol Center, 5 miles west of Canandaigua Lake, contains two or three nonpersistent, very thin shaly, laminated flags which pass into shale along the outcrop. This is the first evidence of the coarser character which dominates the rocks at this stratigraphic level farther east. These are local concentrations of slightly greater amounts of quartz grains in the shale. Some thin flags occur in the shales along the shores of Canandaigua Lake, but their stratigraphic position varies, and here also the layers are not continuous. The upper 15 feet of the West River in this region has been designated a transition series by Clarke and Luther (1904, p. 29) and called the Standish flags and shales. This bed is described as a series of "thin, uneven, bluish gray flags and olive shales". It was originally considered a lentil confined to the Canandaigua quadrangle. It was stated that "the beds show some difference from those below in faunal content". The faunal list given for the Standish by the same authors consists of five species: *Bactrites aciculum*, *Gephyroceras* sp., *Pleurotomaria cognata*, *Pterochaenia fragilis*, and *Ontaria suborbicularis*. Only *Pleurotomaria cognata* is not found in the shales below, although *Ontaria suborbicularis* is confined to the list of "Genundewa" fossils which they separate from the West River. Chadwick (1935, p. 312) lists *Ontaria suborbicularis* as a nondiagnostic form of the pteropod limestone; *Gyroma cognata* (*Pleurotomaria cognata*) is listed as a nondiagnostic form for the Genesee group as a whole. The paleontology does not justify separation of the "Standish" from the West River.

Olive shales, presumably characteristic of the formation according to the original description, are very limited areally even in the type area and are not confined to a single stratigraphic level.

Greenish-brown shale was found at the top of the West River in the stream at Black Point, on the west side of Canandaigua Lake; beds at this level in other near-by streams, however, are dark gray and black like the rest of the West River. Green shale is found, also, close to the base of the formation in one of the streams flowing off Bare Hill, on the east side of the lake, 1.7 miles north of Vine Valley. Five feet of it is interbedded between the shales of the pteropod facies, beginning 11 feet above the base of the first *Styliolina* layer, near the lake shore.

Microscopic examination shows that the green shales differ from the dark layers only in the absence of carbonaceous matter and calcite; the calcite apparently has been leached. Marcasite is present, some of it altered to limonite, and a few specks of hematite were recognized. Occasional dark spots are interpreted as bituminous stains, resulting either from migration of hydrocarbons or from distillation of plant spores in place.

The Standish is not a clearly defined stratigraphic unit in the Canandaigua Lake region, and the flag lenses are not a part of the Starkey tongue. It is recommended that the name Standish be abandoned.

INTERTONGUED STRATA OF THE KEUKA-SENECA-CAYUGA LAKES AREA

General statement.—The interval between the Geneseo shale and the Middlesex shale in the Keuka Lake valley is considerably thicker than at Canandaigua Lake. A tongue of Sherburne type rock in the middle of the West River causes a thickening of the whole sequence to 250 feet. The West River shale is divided into the Penn Yan tongue (lower shales) and the Milo tongue (upper shales) by the penetration of the Starkey tongue of the Sherburne (Fig. 2).

Penn Yan tongue (new).—The Penn Yan tongue is the continuation of the West River shale below the Starkey tongue of the Sherburne. It is named from the outcrops in the streams along the south side of the Keuka Lake outlet, near the town of Penn Yan. It is particularly well exposed in the middle gully of the three draining north into the outlet at Seneca Mills. It can be traced eastward as far as the Cayuga Lake region where it is very thin; farther east it is indistinguishable from the Sherburne. The Penn Yan is approximately 140 feet thick in the Keuka Lake outlet. The base of the tongue is difficult to determine in this area because of the absence of the *Styliolina* facies. Shales in the lower part of the Penn Yan are similar to those of the Geneseo below, but the contact is placed at the base of a series of dark, limy shales and interbedded thin shaly limestone layers which may correlate with the more easily recognized limy layers at the base of the Penn Yan tongue in the Seneca Lake region. This series of interbedded shaly and limy beds occurs approximately 120 feet above the base of the Geneseo shale. The top of the tongue is placed arbitrarily at the base of the Crosby in the Keuka Lake area, although Starkey type flags begin lower.

Lithologically the lower part of the Penn Yan tongue in the type area is similar to the West River in the region to the west. It consists of interbedded black and blue-black shales, blue-gray shales, and thin limy argillaceous beds, all in the medium siltstone size. Fossils are relatively scarce. Common forms include: *Pterochaenia fragilis*, *Buchiola retrostriata*, *Chonetes lepidus*, *Manticoceras sinuosum*, and plant remains. Concretions are plentiful, and some marcasite nodules are present.

Unfossiliferous flags ranging from ½ to 3 inches thick are interbedded in the dark shales beginning 25 to 30 feet from the base. Many of these are cross-laminated. They are coarse siltstones, some with clay dominant and others with quartz. The largest grains are 0.09 mm. in diameter. They are widely spaced in the 50 feet of shale above the first appearance of these coarser layers. They become more closely spaced in the remainder of the Penn Yan and are considerably heavier, reaching 8 inches in thickness. The heavier beds are not ordinarily cross-laminated.

Shales of the upper part of the Penn Yan are generally lighter-colored than those at the base, and contain more fossils. *Paleoneilo constricta*, *Chonetes lepidus*, *Buchiola retrostriata*, *Pterochaenia fragilis*, *Manticoceras sinuosum*, *Bactrites aciculum*, crinoid stems, and trails are common. The pelecypod-rich fauna continues into the shales of the overlying Starkey tongue, so that the passage of the Penn Yan into the Starkey is faunally as well as lithologically transitional.

Starkey type beds enter the section somewhat lower in the Seneca Lake valley than they do to the west, but the main mass of the Starkey begins at approximately the same level, 140 feet above the top of the Geneseo. Medium argillaceous siltstones still predominate but are interbedded with increasing amounts of argillaceous and quartz-rich coarse siltstones. The lower 23 feet of the Penn Yan includes the argillaceous silty limestone facies which occupies the same relative position as the *Styliolina* facies of the West River to the west. Scattered concretions and concretionary layers are present throughout the shales. The top of the Penn Yan tongue is placed at the first occurrence of closely spaced, heavy flags. As in the Keuka Lake area, this is a matter of convenience for field determination and mapping; there is no line of contact between the two tongues, but rather a zone of gradual change by interfingering.

The Penn Yan tongue thins rapidly eastward and pinches out beyond Cayuga Lake. At Trumansburg Creek on the west side of Cayuga Lake it is 36 feet thick. The limy facies at the base consists of 16 feet of interbedded limy argillaceous layers and shales with numerous concretions and is clearly defined in the walls of the stream. The shale of this lower limy facies is predominantly dark to medium gray, weathering to bluish gray. The number and variety of fossils has decreased. Those remaining include *Styliolina fissurella*, *Pterochaenia fragilis*, *Chonetes lepidus*, *Manticoceras* (?) sp., and *Paleoneilo* sp.

The upper strata are soft, gray shales with interbedded black and olive-drab layers. They are slightly micaceous, argillaceous medium siltstones and show in cliffs as a well-marked band of easily weathered rock. Fossils are rare and are the same as those in the limy facies. Despite an apparently sharp contact between the Penn Yan and the Sherburne, there is evidence of intergradation. Flags, abundant in overlying beds, are present throughout the shale of the Penn Yan, but are extremely thin, averaging less than an inch thick. Most of them are cross-laminated.

Milo tongue (*new*).—Shales of the West River above the Starkey in the Keuka Lake valley are called the Milo tongue (Fig. 2) because of their exposure in streams in Milo township. The type section is in Crosby Gully, on the east side of Keuka Lake, 7 miles south of Penn Yan. This stream lies 1¼ miles south of the Milo township line, but it has the finest exposure of the Milo tongue. The Milo is present in

the Keuka Lake valley and in the streams on the west side of the Seneca Lake valley. It grades rapidly into a coarser facies and is not recognized on the east side of Seneca Lake except for relatively thin portions of the West River type shale within the Starkey tongue.

The Milo in the Keuka Lake region is well defined. The base is marked by the Keuka flagstone (Fox, 1932, p. 683), and the top by the first continuous series of black, fissile, bituminous Middlesex shale. Both contacts are gradational, and the lower is placed at the Keuka flagstone for ease of recognition. Fifteen feet of black shale immediately below the Keuka is allied lithologically and possibly belongs with the Milo.

The Milo consists of 90 feet of black and dark-gray fissile shale with interbedded light-gray and blue-gray layers, limy bands, and spherical concretions. Shales are uniformly medium argillaceous siltstones. Concretions are common and range from a few inches to over a foot in diameter. Only a very few thin flags are present. Plant remains are plentiful, and some fragments are several feet long. All are badly crushed and macerated. Faunally the Milo tongue is similar to the shales of the West River to the west.

Sherburne formation.—According to the original definition (Vanuxem, 1842, p. 292), the Sherburne at the type locality near the town of Sherburne, is a sparsely fossiliferous shale and sandstone series overlying the Geneseo shale and terminating at the first appearance in abundance of Ithaca fossils. Cooper and Williams (1935, p. 825–27) indicate that the Sherburne east of the type locality, as identified by workers after Vanuxem, includes beds as low as the Tully, which in this area is of the same facies as the Sherburne.

The Sherburne as used here includes 310 feet of interbedded shales and flags (siltstones and very fine sandstones) between the Penn Yan tongue of the West River and the attenuated remnant of the Middlesex black shale in the Cayuga Lake valley. The term is essentially a lithologic one. The fauna is primarily Naples type, which characterizes the West River, with Ithaca elements. Tracing the Middlesex into this area has given a more reliable method of terminating the Sherburne.

Starkey tongue of Sherburne (new).—A tongue of coarser rock penetrates westward into the West River shale, between the Penn Yann and Milo tongues of the West River in west-central New York (Fig. 2). The term Starkey is used for this unit because of the exposures in Starkey township on the west side of Seneca Lake. The type locality is Crosby Gully, on the east side of Keuka Lake, where the tongue is best exposed, but names available in this area are preoccupied. The term Standish, heretofore used to include the Starkey and Milo tongues, is dropped, because the Starkey does not correlate with the thin irregular lenses originally defined as Standish in the Canandaigua Lake valley (Clarke and Luther, 1904, p. 29–30).

The root of the tongue is the Sherburne formation of the Cayuga Lake area. The Starkey extends westward to within a few miles of the east shore of Canandaigua Lake where it is only a thin finger, approximately in the middle of the West River.

The base of the Starkey in the Keuka Lake region is placed at the Crosby. The top is marked by the Keuka flagstone. The first appearance of heavy and closely

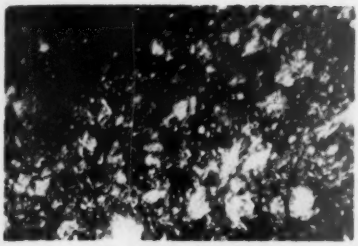


FIGURE 1

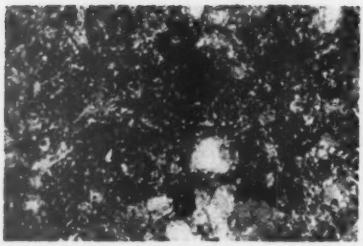


FIGURE 2

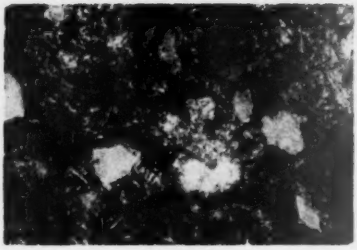


FIGURE 3

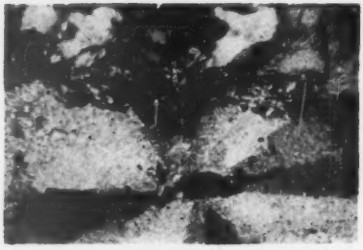
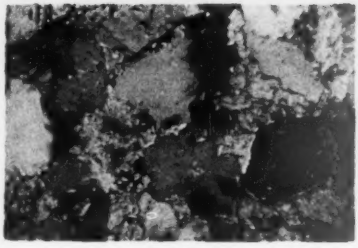


FIGURE 4



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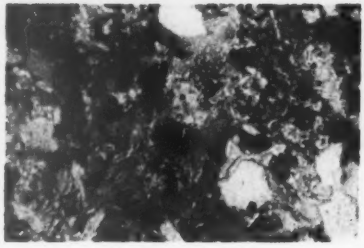

FIGURE 6

FIGURE 1. MEDIUM ARGILLACEOUS SILTSTONE
Top of Geneseo, Lodi Glen, Seneca Lake. Quartz grains more abundant than in Geneseo to west. Mode 0.018 mm. Crossed nicols, X 147.5.

FIGURE 2. MEDIUM ARGILLACEOUS SILTSTONE
Geneseo, 16 feet above base; West Brook, Chenango valley. Fewer quartz fragments than in black shale of Seneca Lake. Considerable carbonaceous material. Mode 0.018 mm. Crossed nicols, X 140.5.

FIGURE 3. COARSE ARGILLACEOUS SILTSTONE
Dark-gray shale in Starkey, 63 feet above base of Crosby lentil, Crosby Gully, Keuka Lake. Similar to West River type rock, but larger quartz grains. Mode 0.03 mm. Crossed nicols, X 140.5.

FIGURE 4. COARSE QUARTZ SILTSTONE WITH CALCITE
Styliolina layer in Starkey, Willow Grove Glen, Keuka Lake. Showing contact between *Styliolina* lens and coarse quartz siltstone above. Radiating calcite below is portion of *Styliolina* shell. Crossed nicols, X 147.5.

FIGURE 5. VERY FINE CALCAREOUS SANDSTONE
Crosby lentil of Starkey, near Seneca Mills, Keuka Lake Outlet. Abundant calcite and unusually large quartz grains. Mode 0.06 mm. Crossed nicols, X 147.5.

FIGURE 6. COARSE QUARTZ SILTSTONE
Keuka lentil at top of Starkey, Crosby Gully, Keuka Lake. Characteristic coarse quartz siltstone of Starkey of this area. Mode 0.039 mm. Crossed nicols, X 147.5.

PHOTOMICROGRAPHS

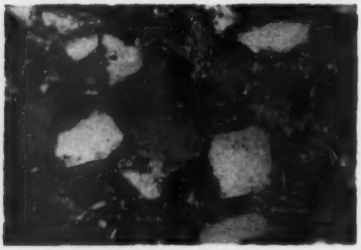

FIGURE 1

FIGURE 2

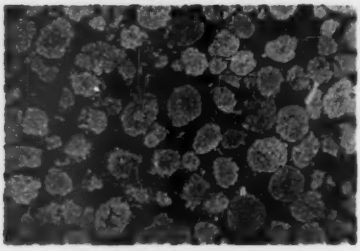

FIGURE 3

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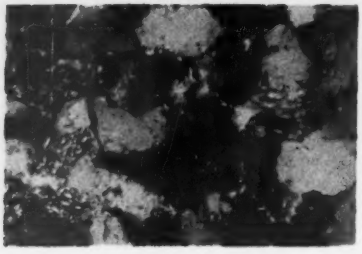

FIGURE 5

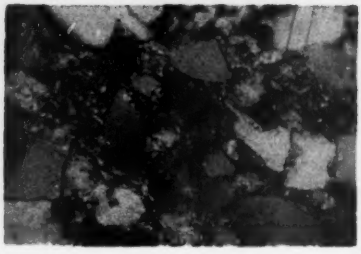

FIGURE 6

FIGURE 1. COARSE QUARTZ SILTSTONE
Starkey, 90 feet from base, Lodi Glen, Seneca Lake. Coarse quartz siltstone approaching very fine sandstone. Example of coarser layers of Starkey of Seneca Lake area. Crossed nicols, X 147.5.

FIGURE 2. COARSE QUARTZ SILTSTONE TO VERY FINE SANDSTONE
Red-brown massive layer, 38 feet below Middlesex, Trumansburg Creek, Cayuga Lake. Showing larger grains which reach 0.08 mm. Mode 0.05 mm. Crossed nicols, X 147.5.

FIGURE 3. MARCASITE PELLETS
Marcasite pellet layer, 14 feet below Plumularia zone of Sherburne, Trumansburg Creek, Cayuga Lake. Showing cracked cores and spherical shapes of pellets and irregular accumulations of marcasite between. Pellets may be replaced plant spores. Reflected light, X 12.

FIGURE 4. COARSE QUARTZ SILTSTONE
Base of Sherburne, Skaneateles Lake. Mode 0.035 mm. Crossed nicols, X 140.5.

FIGURE 5. COARSE QUARTZ SILTSTONE
Near base of Ithaca, Tinkers Falls Ravine. Mode 0.048 mm. Crossed nicols, X 147.5.

FIGURE 6. COARSE QUARTZ SILTSTONE
Unadilla, 162 feet above upper *Hypothyridina* zone of Tully; vicinity of New Lisbon. Mode 0.048 mm. Crossed nicols, X 140.5.

PHOTOMICROGRAPHS

spaced flags identifies the base of the tongue in the area to the east where the Crosby is not recognized; the Starkey continues to the base of the Middlesex in the east side of the Seneca Lake valley and eastward, where the Milo tongue of the West River cannot be defined.

CROSBY LENTIL: The Crosby (Fox, 1932, p. 681) lentil lies at the base of the Starkey in the Keuka-Seneca Lake divide. It is a heavy-bedded crinoidal layer, 6 to 8 feet thick, and is classified as a very fine calcareous sandstone (Pl. 1, fig. 5). Calcite is the predominant constituent in some instances. A few quartz grains reach 0.12 mm. in diameter and exhibit a rough orientation, emphasized by the mica flakes, some of which are bent around quartz fragments. Many of the quartz grains have undulatory extinction and contain minute inclusions, some of which are arranged in more or less regular crisscross pattern. These characteristics conform with those of the Catskill facies for which a metamorphic source is claimed (Mencher, 1939, p. 1770). The same features are common in the grains throughout the Genesee rocks but are better displayed by the Crosby because the individual particles are larger.

Fragments of very fine sandstone size in this area indicate a greater current velocity to transport the heavier particles. The rough orientation of the grains is not necessarily the result of the stronger currents but may be due to slight intraformational movement before consolidation. Other layers such as the Keuka show evidence of this, and the bent mica flakes in the Crosby suggest the same condition.

KEUKA FLAGSTONE: The Keuka flagstone (Fox, 1932, p. 683), at the top of the Starkey, is a 4-inch cross-laminated coarse quartz siltstone with considerable calcite (Pl. 1, fig. 6). It is characterized by an undulatory surface and shows evidence of intraformational contortion. Microscopically it resembles many of the thinner flags of the lower Starkey. It is confined to the Keuka Lake valley and the west shore of Seneca Lake.

Starkey shales and flags.—The Starkey is lithologically like the Sherburne—interbedded medium quartz siltstone flags, gritty micaceous blue-gray and dark-gray shales, and some black shales. The flags are thinner, and many are cross-laminated; quartz grain content decreases westward. Only a few persistent flags remain west of Keuka Lake.

The Starkey is 140 feet thick on the east shore of Keuka Lake. The shales are medium argillaceous siltstones with a few coarse siltstone layers (Pl. 1, fig. 3), and most of the thin flags are the same, with a higher percentage of quartz. The coarser flags are coarse quartz siltstones.

The Starkey is 240 feet thick on the east shore of Seneca Lake, where neither the Crosby nor the Keuka can be distinguished. The thickening results from the change of facies of the Milo tongue of the West River to Starkey type lithology.

A 9-foot series of unfossiliferous heavy flags is the cap rock of the high falls in Lodi Glen on the east side of Seneca Lake. This series is at approximately the same stratigraphic level as the Crosby but otherwise does not resemble it. In this area it is regarded as the base of the Starkey, which includes three limy crinoidal layers similar to the Crosby. These occur at intervals of 65, 85, and 135 feet from the base. The 7-foot middle layer corresponds in thickness to the Crosby but is considerably

higher stratigraphically. The lower and upper calcareous layers are only 16 and 14 inches thick, respectively.

A 12-inch band of black shale, 140 feet above the base of the Starkey, may correlate with the black shale associated with the Keuka flagstone to the west. Shales interbedded between the flags above the black shale band are more micaceous and gritty than those of the Milo tongue in the Keuka Lake valley. No real distinction is possible between the rocks above and below the black shale horizon.

The close association of the highest calcareous, crinoidal bed with the 12-inch black shale band prompted Clarke and Luther (1909, p. 29) to designate them erroneously as Parrish limestone and Rhinestreet shale, respectively. Luther found these limestone and shale beds only on the west side of Seneca Lake in Plum Creek, but they are well exposed in Mill Creek which flows into Lodi Glen on the east slope of the Seneca Lake valley.

The western extremity of vigorous current action on the marginal marine plain of the Sherburne and Starkey is marked approximately by the east shore of Seneca Lake. The Starkey of this area has more flaggy layers (coarse siltstones for the most part) than that of the Keuka Lake valley (Pl. 2, fig. 1), and many of the shaly beds contain quartz grains of coarse siltstone size. The velocity of the currents probably was controlled by the depth of the water. In addition to larger fragments, effective current action is indicated by channel filling, cross-bedding, and other sedimentary structures. The limiting depth of vigorous current action off the New England coast is approximately 200 feet (Stetson, 1938, p. 34). Although the situation is not precisely comparable, it provides some basis for the determination of the depth of water on the Genesee marginal marine plain. The depth of water in which the Sherburne and the eastern portion of the Starkey were deposited probably ranged from a few feet to 200 feet. Starkey beds to the west contain smaller particles on the whole, carried by currents reduced in velocity because of the increased depth.

The contact between the Starkey and overlying Middlesex is distinct in Mill Creek. The Middlesex thins to about 27 feet, and only the lower 10–15 feet is densely black shale characteristic of the formation in western New York. A change of facies is noted in the upper part where the shales are interbedded with thin flags and finally merge into the overlying strata.

Fossils are sporadic through the Starkey beds. Calcareous layers are characterized by an abundance of the coral *Cladochonus* in addition to crinoid fragments. *Manticoceras* sp., *Leiorhynchus quadricostatum*, and *Chonetes lepidus* occur as well as the small Naples type faunule which characterizes the West River. On the west side of the lake, Clarke and Luther (1909, p. 30) found Ithaca elements in a limestone layer which corresponds in stratigraphic level to the highest limestone at Mill Creek. This bed contains *Leptostrophia mucronata*, *Microspirifer mucronatus* var. *posterus*, *Reticularia laevis*, *Cladochonus*, and crinoid stems.

Flags and shales in the upper part of the formation are apparently even less fossiliferous than those below. They contain a mixed Naples-Ithaca faunule including: *Leiorhynchus quadricostatum*, *Chonetes lepidus*, *Atrypa reticularis*, *Productella spinulicosta*, *Pterochaenia fragilis*, *Ambocoelia umbonata*, *Buchiola retrostriata*, *Paracardium doris*, *Honeoyea major* (?), *Bactrites* sp., *Manticoceras sinuosum*, *Probeloceras lutheri*, and *Styliolina fissurella*.

Sherburne of Cayuga Lake.—The Sherburne is well exposed in Trumansburg Creek, on the west side of Cayuga Lake. It consists of interbedded dark-gray, gritty, micaceous shales and medium to coarse quartz siltstone flags which range from 1 inch to 2 feet thick. The shales are somewhat coarser than those of the West River, owing to the greater percentage of quartz grains and the somewhat larger mica flakes. They are sparingly fossiliferous. Cross-laminated, unfossiliferous flags are more closely spaced and thicker in the middle and upper portions, though a series of resistant and closely spaced flags about 50 feet from the base has formed the cap rock for the high falls in Trumansburg and Taughannock creeks. Siltstone layers in the interval between 100 and 200 feet above the base of the formation tend to pinch out rapidly. Associated with these are beds showing intraformational contortion. Channel filling and cross-bedding indicate strong current action.

At the top of the second falls in Trumansburg Creek there is an accumulation of marcasite pellets, some of which may be replaced plant spores. A great abundance of badly macerated plant remains form a considerable part of the beds immediately adjacent. Powdery yellow, black, white, and brown sulphides coat the weathered surfaces of these layers.

Many of the irregular siltstone lenses are reddish brown even on fresh surfaces and could be called red beds. They are coarse quartz siltstones (Pl. 2, fig. 2) having over 50 per cent quartz, with modes ranging from 0.036 mm. to 0.048 mm., and as such do not differ from the more persistent flags. The largest grains observed have a diameter of 0.1 mm. Some of these red beds merge into gray beds along the outcrop. Other flags have reddish surfaces but are gray in the inner portions. The red color penetrates as much as an inch in some of these layers and appears to result from weathering. Coloring matter seems to be a thin coating of limonite on the quartz grains. The limonite may be altered from marcasite.

Ten feet of the Sherburne immediately below the Middlesex consists of soft, dark-gray shale which weathers into small blocks. This shale is less resistant than that below and may be part of the Middlesex. For convenience, however, the base of the Middlesex is placed at the first appearance of densely black, fissile shale approximately 300 feet above the top of the West River. The change in facies in the Middlesex is nearly complete at this locality.

The lower 150 feet of the Sherburne carries a Naples fauna. Poor preservation and difficulty in breaking out individuals from the rock are in part responsible for the small list of forms which includes: *Styliolina fissurella*, *Pterochaenia fragilis*, *Buchiola retrostriata*, *Paracardium doris*, *Paleoneilo constricta*, *Lingula spatulata*, *Chonetes lepidus*, *Bactrites aciculum*, *Probeloceras lutheri*, *Grammysia subarcuata*, *Leiorhynchus quadricostatum*, *L. mesacostalis*, *Cladochonus* sp., plant remains, and crinoid stems.

Shales approximately 160–170 feet from the base of the Sherburne in Trumansburg Creek contain abundant fernlike remains which have been referred to as *Lycopodites vanuxemi* and *Plumalina plumaria*. Similar forms occur in the Losh Run of Pennsylvania. These are thought to be hydroids, resembling the modern hydrozoan *Plumularia*. Associated with this zone are abundant *Leiorhynchus quadricostatum*, *Chonetes lepidus*, and some poorly preserved crinoids. It is doubtful that this is the same "*Lycopodites*" zone (Williams, 1909, p. 8) classified as lower Ithaca.

Fossils are scarce in the rocks above this zone and are similar to those below.

Ithaca-type productids are abundant in the dark shale about 60 feet above the Middlesex.

Sherburne east of Cayuga Lake.—The Sherburne east of Cayuga Lake, including the Moravia, Skaneateles, and Tinkers Falls sections, consists of coarse argillaceous, shaly, and flaggy siltstones. Interbedded coarse quartz siltstones (Pl. 2, figs. 4, 5) increase in amount eastward. Although medium argillaceous siltstones are present throughout these beds, they are more numerous lower in the sections.

The Sherburne at De Ruyter is apparently similar to that to the west. The lower portions of the overlying Otselic, however, has very fine quartz sandstones.

The Onteora formation of the Catskills, approximate equivalent of the Sherburne, contains graywackes, the larger grains of which are 0.5 mm. or larger, with quartz in some specimens constituting 80 per cent of the rock (Mencher, 1939, p. 1774). The size of the mode of these grains is undoubtedly smaller, but in any case the facies shifts eastward again into a larger size grade with increased quartz.

Stratigraphic divisions east of the Cayuga Lake area do not correspond with those to the west. In the Skaneateles Lake region, the top of the Sherburne is placed at the *Reticularia laevis* zone or Cornell member (Smith, 1935, p. 58, 61–62), 220 feet above the top of the Geneseo. The base of the Sherburne is lower than at Cayuga Lake because it includes some of the upper Geneseo which has changed facies; thus, a point 220 feet above the top of the black shale in the Skaneateles area will be stratigraphically below a horizon measured on the same basis to the west (Fig. 4). The top of the Sherburne projected from Cayuga Lake would be more than 100 feet higher than the Cornell member. Inasmuch as the beds above the Cornell member have Ithaca fossils and belong to the Ithaca "formation", the Ithaca is in part equivalent to the Genesee group even this far west.

Variable occurrence of the *Reticularia laevis* zone is indicated by the fact that it apparently is not persistent and in places is below and above the horizon of the Cornell member. For the Skaneateles region, Smith states that "the attempt is made to adhere to this practice [of placing the base of the Ithaca above the *Reticularia laevis* zone], although lack of data has frequently forced an arbitrary placing of the line" (Smith, 1935, p. 62). Furthermore, he discovered another *Reticularia laevis* zone about 315 feet above the Geneseo and was forced to give it "a tentative position in the Ithaca member" (1935, p. 63). A more reliable guide is essential, but until one is discovered, the present unsatisfactory method of classification of the Sherburne in this area is retained.

Farther east the method of delineation of the Sherburne is even less satisfactory, since it depends on the first appearance of abundant Ithaca fossils. Such a designation is at best general, and the top of the Sherburne in any given locality will depend on chance occurrence and preservation of fossils and the definition of an "abundance". Ithaca forms, furthermore, come into the sections stratigraphically lower toward the east. It may be necessary to abandon the term Sherburne in this area and include the strata in a larger, better-defined unit.

SUMMARY OF CONDITIONS OF DEPOSITION

The presence in the Genesee silty facies (Starkey-Sherburne) of ripple marks, cross-lamination, clay galls, lensing of beds, and channel-filling indicates shallow-water

deposition, probably on a marginal marine plain which graded eastward into deltal (Barrell, 1913, p. 429–472; 1914, p. 87–109, 225–253) and alluvial (Mencher, 1939, p. 1783) deposits. The large areal extent of these sediments and the abundance of plant remains, worm burrows, animal trails and tracks, drag markings, and ice-crystal impressions support this view.

Gradation and interfingering of black shales into coarser beds to the east seems significant. Quartz grains in the black shales are only slightly smaller than those in the rocks to the east, which were deposited in water probably less than 200 feet deep. No abrupt change in grain size occurs, such as might be expected if shallow currents flowing over a marginal marine plain suddenly debouched into deep water. The decrease westward in amount of quartz is also gradual, and the relative scarcity of quartz grains in the western black shales reflects distance from the source rather than abrupt deepening of the basin. The black shales of the Genesee, though deposited in deeper water than the Starkey and Sherburne, probably were laid down in water only 200 or 300 feet deep. Finely laminated black shales indicate very quiet conditions of deposition, and high carbonaceous and bituminous content plus widely disseminated iron sulphides suggest stagnation.

The Starkey tongue represents a slight marine regression with extension of the marginal marine plain. The north-south boundaries of the tongue are not known. To the north the Genesee beds are eroded away and available information does not indicate a tongue at the same stratigraphic level in Pennsylvania. Carefully logged wells in the intervening area, where the Genesee strata are covered, might show the southern extent and form of the Starkey, but these data are at present insufficient.

The cause of the marine regression cannot be ascertained. Epeirogenic uplift is more probable than eustatic withdrawal, since minor variations in intensity of the orogeny affecting the highland is to be expected. The simplest explanation for the presence of the tongue is a shift in the direction of the main streams flowing on the delta, unrelated to epeirogenic or eustatic movement.

The climate of the early Upper Devonian in this area seems to have been little different from that of the present. Seasonal changes of a temperate climate account for the presence of ice and corals in the same rock series.

The Upper Devonian beds form part of a flysch which indicates a progressively rising land mass to the east. The Genesee beds and their equivalents bear out the evidence of the rocks in the Catskills which "convey clearly the picture of an orogeny of not very great intensity beginning early in Hamilton time . . . and gradually increasing slightly in intensity throughout the Middle and Upper Devonian with the maximum development in the late Upper Devonian and probably after the highest of the present New York Catskill beds had been laid down" (Mencher, 1939, p. 1785).

CORRELATION

Correlation of Genesee equivalents east of Cayuga Lake is largely conjectural. No known adequate faunal or stratigraphic horizon carries lines of contemporaneity eastward into and through the strata of this area. Projecting the top of the Genesee group into the area where there is no stratigraphic evidence to continue such a line is valueless as an aid in mapping formations and structures. It will give a clearer con-

ception, however, of the approximate portion of the mass of beds in eastern New York which are equivalent to the Genesee group.

The rate of thickening of the Genesee beds from Lake Erie to Cayuga Lake is roughly 4 feet per mile. The rate varies, however, in smaller sectors within that area. Between Lake Erie and Canandaigua Lake the rate is relatively uniform at about 2.8 feet per mile. A sudden increase to 13 feet per mile between Canandaigua Lake and Keuka Lake is due to the penetration of the Starkey tongue. The thickness of the group between Keuka Lake and Cayuga Lake is constant; it is not likely, however, that the rocks corresponding in time to the Genesee group will maintain an equal thickness as the source is approached. Hamilton and Tully beds thicken to the east, and probably the overlying rocks do the same. That the average rate of thickening of 4 feet per mile from Lake Erie to Cayuga Lake will be maintained eastward is only supposition. Though admittedly unreliable, it is the only figure available for use in such a projection.

Adoption of the rate of thickening of 4 feet per mile places the projected top of the Genesee group (Fig. 4) approximately 550 feet above the base of the Geneseo in Moravia, 580 feet at Skaneateles Lake, 620 feet at Tully, 660 feet in the vicinity of De Ruyter, 740 feet in the Chanango valley, 780 feet at New Lisbon, and more than 1000 feet in the Catskill front.

The Ithaca is divided into the Otselic member below and the Cincinnatus member by Clarke (1903, p. 24) from the vicinity of De Ruyter eastward to the Chanango valley. The projected Genesee group includes the beds to the top, or nearly so, of the Otselic division in this area. Still farther eastward the Unadilla and part of the Oneonta formation are included. Beyond the Otego valley the term Gilboa has been applied to the beds from the top of the Hamilton to the base of the red beds. Since the top of the Tully is not recognized this far east, it is impossible to indicate how much, if any, of the Gilboa is equivalent to the Genesee group. Similarly the thickness of red beds over the Gilboa assignable to the Genesee cannot be ascertained. The Genesee equivalents in the Catskill area probably occur above the Onteora formation of Chadwick. This assumption is based on the equivalence of the Onteora beds to the Gilboa (Cooper and Williams, 1935, p. 819).

Correlation of the New York Genesee with the beds in Pennsylvania is hazardous. Willard (1939, p. 238–239) has attempted such a correlation. Some refinements are possible after comparing Willard's data and those from the present work. The Geneseo shale of New York correlates faunally and stratigraphically with the Burkett shale of Pennsylvania. Examination of the faunal lists of the Harrell shale indicates a close relationship with that of the West River of New York. The Starkey tongue of the Sherburne and the Sherburne itself seem to be equivalent to the Losh Run and part of the Brallier. The trail *Pteridichnites biseriatus*, used as a guide fossil for the Brallier by Willard, is found in the Penn Yan tongue, the Starkey, and the Sherburne of New York. It is a facies fossil, however, and of doubtful significance.

It is not possible to correlate adequately between the Genesee beds of west and central New York and their eastern equivalents, so correlation of the Genesee rocks of eastern New York with those of Pennsylvania is limited to generalizations. It seems likely, nevertheless, that the eastward portion of the New York Sherburne, including the Geneseo equivalents, is equal in part to the Trimmers Rock of Pennsylvania.

SUMMARY AND CONCLUSIONS

The eastward increasing coarseness of the Genesee group is due primarily to an increasing percentage of quartz grains rather than a spectacular increase in size of grain. The increase in percentage of quartz fragments precedes the shift to larger sizes in the beds where the facies is undergoing gradual change. In general, the change of facies from west to east, as far as the Sherburne valley, involves a shift of only one size grade, from medium siltstone to coarse siltstone. The change is gradual and of an interfingering nature, indicating a steadily diminishing velocity of the currents carrying the particles. It is likely, therefore, that the currents flowed into water only slightly deeper than that of the marginal marine plain represented by the Sherburne and Starkey, and that the black shales are relatively shallow water deposits.

The Genesee group does not fit into the simple sequence of magnafacies described for the Upper Devonian rocks of Pennsylvania. The many variations in the New York strata make it inadvisable to adopt that classification for the Genesee beds.

The isolithic boundary, connecting the areas of change from black shales to flags and gray micaceous shale in New York and northern Pennsylvania, must be extended southward to other localities before it will be useful in indicating the trend of the highlands supplying the sediments.

The Geneseo is regarded here as an integral part of the New York Upper Devonian. Paleontologic and lithologic evidence and the field relationships indicate separation of the Tully from the Geneseo by an unconformity of greater importance than that between the Tully and Hamilton in the area under discussion. Though this local evidence suggests the advantage of placing a principal stratigraphic limit at the base of the Geneseo, the classification of Tully and Geneseo as Middle and Upper Devonian is dependent on their precise correlation with the Givetian and Frasnian of Europe.

The *Styliolina* beds, formerly known as Genundewa limestone, are a series of lentils with highly irregular occurrence and unpredictable thickness at any given locality. They should be regarded as a facies of the West River rather than as a separate formation. Argillaceous limestone layers at the base of the Penn Yan tongue are another facies of the West River.

The Standish is not recognized as a stratigraphic unit distinct from the West River in the type area; the term is abandoned. Lenses originally defined as Standish do not correlate with the Starkey and Milo tongues which erroneously have been called Standish.

The West River formation is split into two tongues in the area from Canandaigua Lake to Cayuga Lake by the Starkey tongue of the Sherburne penetrating from the east. Shales below the Starkey are named the Penn Yan tongue; those above are the Milo tongue.

The presence of the Starkey tongue is explained by a slight marine regression causing a temporary extension of the marginal marine plain in Sherburne time.

The Genesee group, for all practical purposes, terminates in the Cayuga Lake valley. The Sherburne of Cayuga Lake occupies the interval between the top of the Penn Yan tongue and the base of the attenuated remnant of the Middlesex. The top of the Sherburne formation east of Cayuga Lake is defined by the first appearance

of Ithaca fossils, causing an unnatural thinning of the formation because these facies fossils drop stratigraphically to the eastward. For purposes of general correlation in the area east of the Cayuga Lake region, it is necessary to project known horizons from the west because of the absence of other criteria.

Present methods of classification of Upper Devonian strata in east-central New York are unreliable. If no better means of correlation are found, it will be necessary to abandon many of the lesser stratigraphic divisions which cannot be substantiated and use only larger mappable units which are well defined.

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STRUCTURE AND PETROLOGY OF SPECIMEN MOUNTAIN, COLORADO

BY ERNEST E. WAHLSTROM

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ABSTRACT

Specimen Mountain, an eroded Cenozoic volcano in the northwestern corner of Rocky Mountain National Park, Colorado, is flanked on the south and west by an extensive series of faulted and folded lava flows and pyroclastics. Regional studies indicate that the erupted materials spread

laterally for several miles in all directions from a single central conduit and buried a prevolcanic surface of considerable relief.

The volcanic rocks, as shown by analyses, are quartz latite and silicic rhyolite. Quartz latite flows are chilled samples of a magma which by gravitative differentiation gave rise to silicic rhyolite occurring as flows and plug rocks. Thick beds of ash, agglomerate, and breccia attest to intermittent explosions of great magnitude.

Inconclusive evidence suggests a Miocene or Pliocene age for the volcanics.

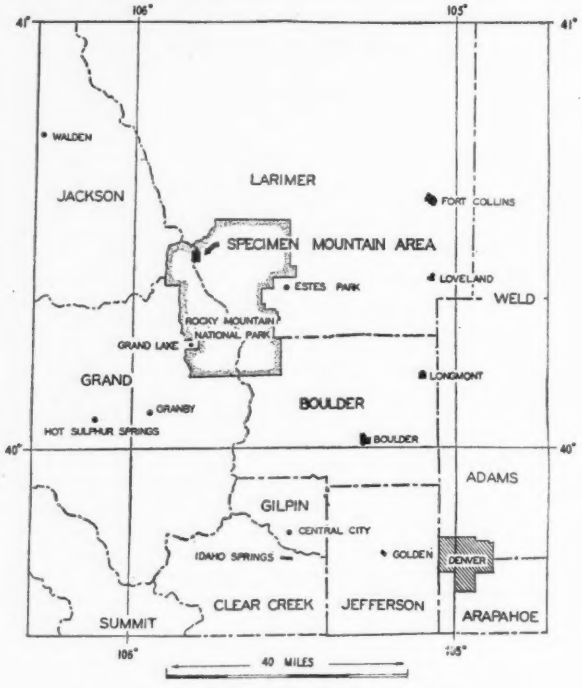

FIGURE 1. *Outline map of north-central Colorado*
Showing location of Specimen Mountain area.

INTRODUCTION

GENERAL STATEMENT

The study of Specimen Mountain was undertaken in collaboration with the National Park Service in an effort to unravel the geological history of this prominent and often-visited landmark. The project is one of several leading ultimately to the reconstruction of the geologic and physiographic history of the whole of Rocky Mountain National Park. This paper summarizes results of geologic and petrologic investigations begun in the summer of 1941 and continued intermittently through the winter of 1942–1943.

LOCATION

Specimen Mountain straddles the Continental Divide at the crest of the Front Range of Colorado (Fig. 1). Its western slopes drain into the headwaters of the

Colorado River in the northwestern corner of Rocky Mountain National Park. The Cache Poudre River, a tributary of the South Platte River, drains the southern and eastern slopes.

The mountain is easily reached by the Trail Ridge Road from either Grand Lake or Estes Park. A trail starting at Poudre Lake leads to the top of Specimen Mountain.

TOPOGRAPHY AND PHYSIOGRAPHY

The elevation of the summit of Specimen Mountain is 12,482 feet. Slopes on the northwest and south sides of the mountain drop steeply to the floors of canyons 2500 to 3000 feet below the summit. The area surrounding Specimen Mountain is exceedingly rugged and displays all the depositional and erosional features of intense alpine glaciation. Specimen Mountain itself is glaciated and is deeply embayed on its northern and western slopes by cliff-walled cirques, which are rapidly filling because of landsliding in the rapidly disintegrating volcanic rocks.

Evidences of frost action and solifluxion are abundant on the upper exposed slopes. Polygonal rock patterns and rock streams are locally conspicuous.

In the pre-Cambrian rocks near Specimen Mountain much of the surface has a "grain" which reflects the gneissoid structure of the underlying rocks. Lenticular masses of pegmatite and granite strike northeast and, in general, stand out in ridges formed by more rapid erosion of adjacent layers of schist and gneiss.

METHODS OF MAPPING

Geologic mapping was done on aerial photographs. A composite geologic map was made from the various photographs, and errors due to tilting and perspective were adjusted by use of a triangulation net surveyed with a plane table and alidade.

The geologic map was then superimposed on a photographic enlargement of a portion of the United States Geological Survey's topographic map of the Rocky Mountain National Park Quadrangle. Because of the inaccuracy of the enlarged topographic map only the 500-foot contours are shown in the finished maps.

ACKNOWLEDGMENTS

The writer wishes to express his appreciation of the assistance given in the field by Mr. Olwin Woodbury. The work was considerably expedited by the helpful cooperation of several officials of Rocky Mountain National Park. The writer is especially indebted to Mr. David H. Canfield, Park Superintendent, Mr. Raymond Gregg, Park Naturalist, and Dr. Carroll H. Wegemann, Regional Geologist, for many courtesies. H. E. Rothrock, Assistant Chief of the Naturalist Division of the Park Service, assisted in many ways. The Council on Research of the Graduate School of the University of Colorado generously provided grants partly to pay for the field work.

GENERAL GEOLOGY

Specimen Mountain is a deeply eroded Cenozoic volcano. Much of its history is revealed in a thick series of pyroclastics and flows. Extensive beds of volcanic ash, breccia, and agglomerate record several periods of major explosive activity (Pl. 1).

The volcano probably developed a normal "Vulcano type" of cone consisting of

alternating surface flows and pyroclastics dipping away on all sides from a single central conduit. The cone was somewhat asymmetric in its early stages because the relief of the surface on which the effusive materials were laid down was considerable. As the cone grew, the irregularities of the prevolcanic surface were buried, and in its later stages the volcano formed a high relatively symmetrical cone.

Many of the extensive flows and pyroclastics exposed on the west and south sides of Specimen Mountain were included in downfaulted and folded blocks which were protected from erosion through much of its post-volcanic history. Denudation has removed the flanking effusive materials on the north and east sides.

The Specimen Mountain volcano emitted two chemically contrasted rock types. The older rocks are quartz latite ash and lava flows. The quartz latite flows rest on the ash or directly on the eroded surface of pre-Cambrian rocks. For the purposes of description these older flows are called the "basal flows".

Above the "basal flows" and apparently formed by differentiation from the quartz latite magma is a thick series of predominantly pyroclastic rhyolitic rocks. Among the last effusive rocks to be erupted were the "upper flows", which have the same composition as the rocks in the plug filling the conduit. The upper flows are several hundred feet thick and locally contain interbedded lenses of ash and breccia.

STRUCTURE

FAULTING AND FOLDING

Four major faults and several minor faults cut the effusives and plug rocks (Figs. 2, 3). The fault planes are curved and dip steeply to the north and northeast at 45° to 80°; they trend on the average northwest-southeast. The faults did not develop simultaneously, for three southeasterly faults are cut and slightly offset by a fourth fault of easterly trend (Fig. 2). Late faults of small displacement intersect and slightly offset the larger faults.

The bedded rocks and flows adjacent to the fault planes are drag-folded considerably out of harmony with the original simple structure. The beds and flows under the lowest and most westerly fault dip about 15° into the plug instead of away from it.

Between the lower and upper faults the layers show all dips from horizontal to vertical. Rhyolite flows near the top of the cliff on the west face of Specimen Mountain are vertical.

STRUCTURAL HISTORY

A series of idealized east-west cross sections (Fig. 4) show the probable structural development of Specimen Mountain.

The oldest ejected material is quartz latitic ash which was thrown out on and in part blanketed a region of considerable relief. Next, quartz latite porphyry flows were erupted and further subdued the relief. In the early stages of volcanism the conduit was probably small (Fig. 4A).

After a period of quiescence following the eruption of the quartz latite flows, major explosions produced a thick series of rhyolitic ash, tuffs, agglomerates, and breccias. Great quantities of the older flows and pre-Cambrian gneiss and granite were included as boulders in the agglomerate and were torn from the walls of a rapidly widening conduit (Fig. 4B).

Showing cliff exposures of lava flows and pyroclastics.

WEST SIDE OF SPECIMEN MOUNTAIN

FIGURE 1. AGGLOMERATE
Southwest side of Specimen Mountain, near the "Saddle". Light gray slope (A) in lower background is volcanic ash. Courtesy H. G. Rodeck.

FIGURE 2. CLIFF EXPOSURE OF VOLCANIC ROCKS
Just west of "Saddle".
A. Ash and breccia with agglomerate layers; B. Pitchstone; C. Ash and breccia; D. Pitchstone; E. Dense rhyolite; F. Ash; G. Mud flow.

The period of major explosive activity terminated with large-scale eruption of rhyolite flows (Fig. 4C). These probably continued until the weight of the flows and pyroclastics in the cone became so great that the western part of the cone collapsed

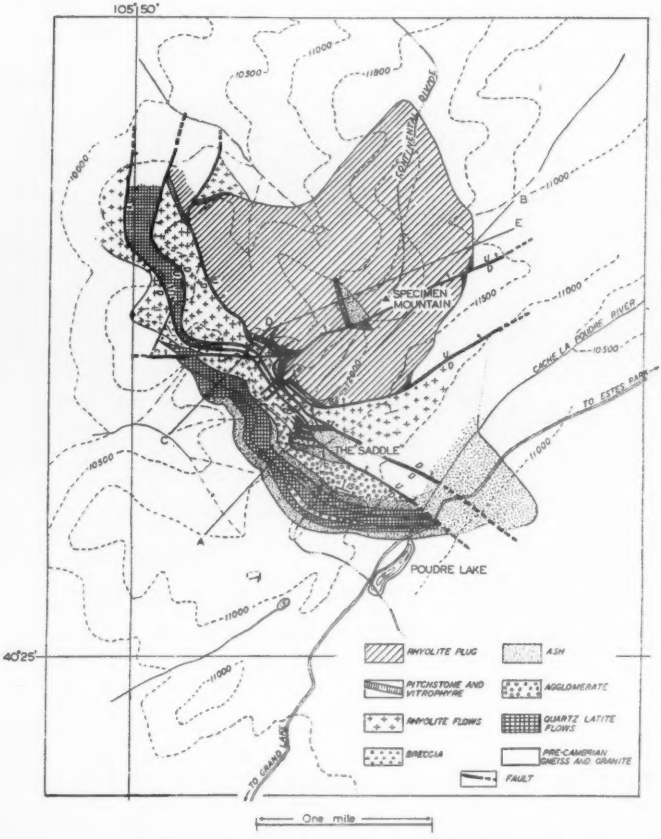

FIGURE 2. *Geologic map of Specimen Mountain area*

(Fig. 4D). Downfaulting and folding may have been simultaneous in part with the last eruptions, although the fact that the faults cut the plug rocks indicates faulting after solidification of the rock, at least in the upper portions of the conduit. Extensive hydrothermal alteration and deposition along the fault planes suggest that the rocks at depth were still hot or even molten at the time of faulting. Probably the downfaulted blocks caused concomitant lateral migration of molten rock in the magma chamber.

The final stage in the history of the Specimen Mountain volcano was the erosion of the bulk of the volcanics and the formation of the present erosion surface (Fig. 4E).

REGIONAL RELATIONSHIPS

Spock (1928) in a reconnaissance survey which included the Never Summer Range west of Specimen Mountain noted extensive capping flows and intrusions ranging from basalt to rhyolite. His report is chiefly petrographic. Many of the flow rocks

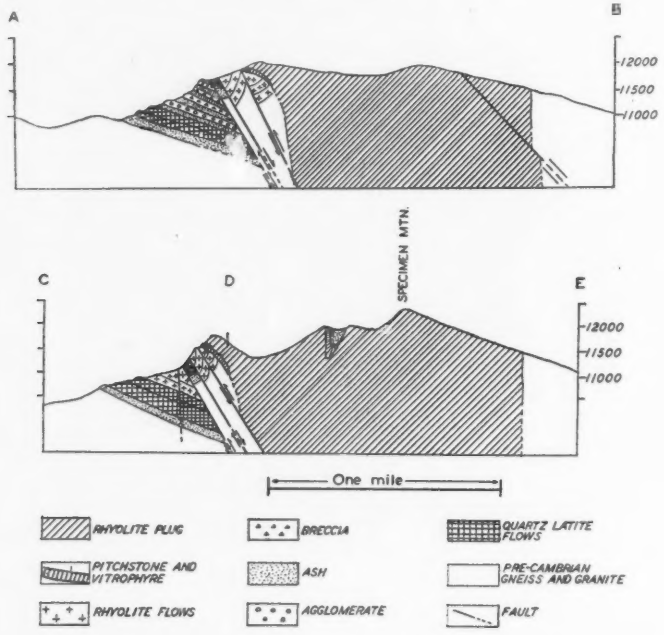

FIGURE 3. *Geologic cross sections of Specimen Mountain*

he describes are virtually identical with the Specimen Mountain rocks, and he points out that although he found no vents for the flows in the Never Summer Range they probably came from the east. The likely source is the vent that supplied the Specimen Mountain volcanics.

Rhyolite flows from Specimen Mountain extended southeast for at least 3 miles. An isolated remnant of rhyolitic flow rock is exposed in the cliffs of the cirque containing Iceberg Lake (Fig. 5). This flow filled a southeasterly trending valley cut into pre-Cambrian gneiss and granite.

The surface on which the flow rests has an elevation of about 11,500 feet and is only about 700 feet lower than the present top of Specimen Mountain. Projection indicates that the top of the volcanic cone of the ancestral Specimen Mountain must have been at least 1000 feet above the top of the present mountain. This conclusion ignores any effects of crustal deformation on the differential levels of the vent and its outermost flows and assumes that the slopes of the sides of the cone exceeded 5°.

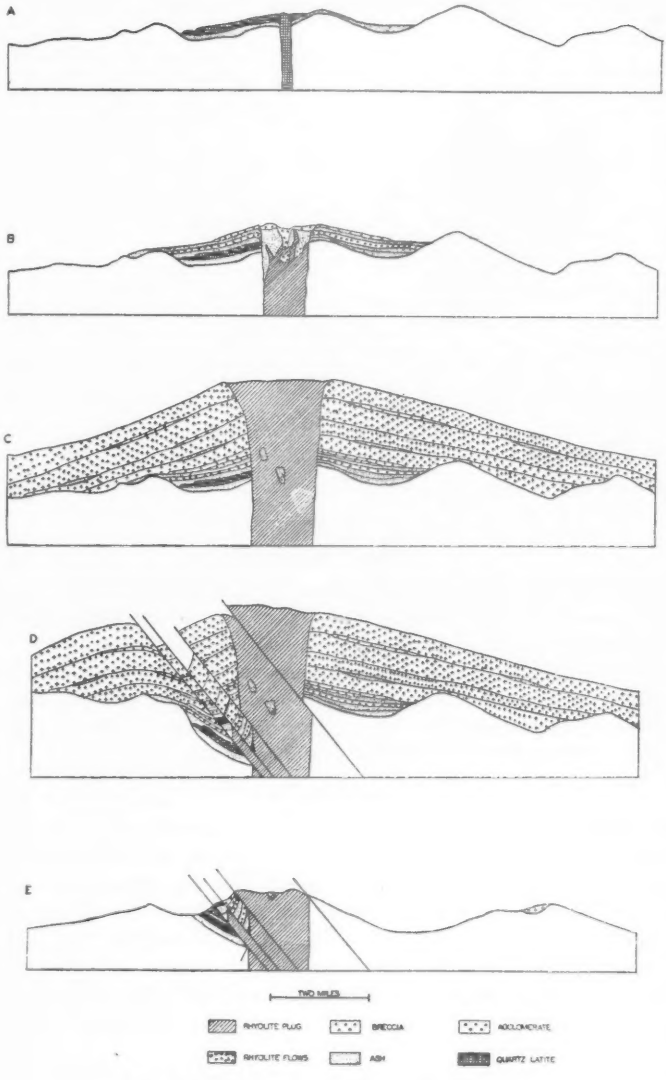

FIGURE 4. *Idealized cross sections*

Showing stages in structural history of Specimen Mountain.

A. Initial eruption of quartz latite ash and flows.
B. Explosive eruption of ash, agglomerate, and breccia.
C. Eruption of rhyolite flows.
D. Collapse faulting and down folding.
E. Erosion to present surface.

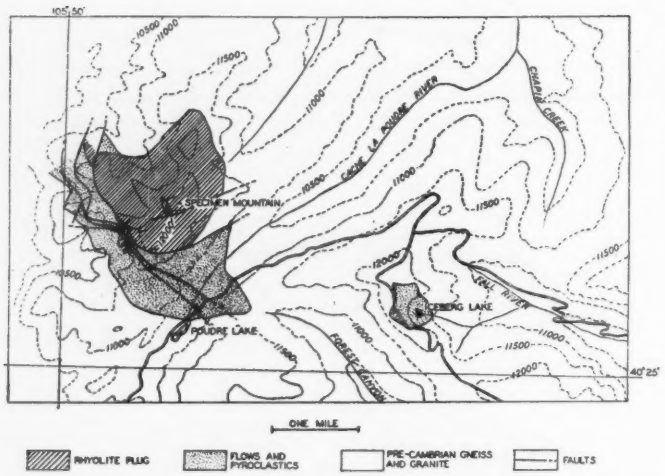

FIGURE 5. *Geologic map showing relationship of flow at Iceberg Lake to Specimen Mountain*

PETROGRAPHY

GENERAL STATEMENT

In order chemically to classify and compare the various rocks in Specimen Mountain the author made 14 quantitative analyses. The analyses were made according to the procedures for rock analyses outlined by Washington (1930).

Chemical analyses of the chief types of igneous materials are presented in Table 1. Inasmuch as the water content of the rocks shows a considerable range, the analyses were recomputed to a water-free basis to bring out more clearly the similarities and differences. Norms were calculated from the analyses.

BASAL FLOWS

Fresh exposures of the basal flows are dark gray or greenish gray. Surface alteration intensifies the green color or produces purple, brown, and orange colorations.

Visible in hand specimen are phenocrysts of plagioclase, needles of black amphibole, and plates of black biotite, all embedded in a fine-grained holocrystalline groundmass.

Microscopically the plagioclases is zoned euhedral labradorite ranging from Ab_{55} at the core to Ab_{45} in the rims. Basaltic hornblende, biotite, and rare augite occur in euhedral crystals. Resorption of both hornblende and biotite has produced clouds of secondary magnetite crystals. Primary magnetite and apatite are scattered sparsely through the rock.

The groundmass makes up about 60 per cent of the rock and contains tiny interlocking rectangular crystals of alkali feldspar and, probably, quartz. The exact mineral content of the groundmass could not be ascertained, but a calculation of the norm indicates that the groundmass must contain abundant potash feldspar and quartz.

From thin-section studies and chemical analyses the rock is classified as a calcic

TABLE 1.—*Rock analyses, Specimen Mountain*

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
SiO_2	63.18	63.45	72.06	77.59	75.44	77.01	68.48	75.35	68.94	76.61	70.00	68.93	77.83	77.35
TiO_2	1.22	1.06	0.27	Trace	0.08	Trace	0.36	Trace	0.25	0.22	0.21	0.34	0.07	0.12
Al_2O_3	17.30	15.89	13.55	12.45	11.97	12.60	12.58	13.31	14.35	12.00	15.46	15.69	11.71	11.83
Fe_2O_3	4.83	3.68	2.14	0.85	1.07	0.48	2.29	1.24	1.26	1.29	2.27	1.56	0.88	1.11
FeO	0.86	1.89	0.67	0.25	0.31	0.40	0.27	0.19	0.95	0.36	0.12	0.18	0.37	0.48
CaO	2.63	3.97	0.27	0.19	0.54	0.12	1.35	0.36	1.30	0.98	0.89	1.18	0.17	0.43
MgO	0.82	0.77	0.06	0.03	0.06	0.03	0.04	0.04	0.08	0.09	0.09	0.15	0.03	0.19
Na_2O	3.68	3.76	2.80	4.23	3.00	4.43	2.57	3.95	3.64	3.46	3.63	4.66	3.31	3.11
K_2O	4.38	3.70	4.55	4.57	5.14	4.35	4.80	5.00	4.69	4.20	6.58	6.35	4.67	4.65
H_2O	1.19	1.41	3.15	0.35	3.00	0.18	7.75	0.39	4.52	1.20	1.15	0.51	0.57	1.30
CO_2	None	0.17	None	None	None	None	None	None	None	None	None	None	None	None
Total	100.09	99.75	99.52	100.51	100.61	99.60	100.49	99.83	99.98	100.41	100.40	99.55	99.61	100.57

ROCK ANALYSES, WATER-FREE

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
SiO_2	63.89	64.55	74.77	77.47	77.30	77.46	73.86	75.77	72.22	77.22	70.53	69.61	78.60	77.93
TiO_2	1.23	1.07	0.28	Trace	0.09	Trace	0.38	Trace	0.26	0.22	0.21	0.35	0.07	0.12
Al_2O_3	17.50	16.14	14.08	12.42	12.10	12.68	13.57	13.40	15.04	12.10	15.58	15.84	11.82	11.92
Fe_2O_3	4.88	3.74	2.22	0.85	1.10	0.48	2.47	1.24	1.32	1.30	2.28	1.57	0.89	1.12
FeO	0.87	1.92	0.68	0.25	0.33	0.40	0.29	0.19	0.99	0.36	0.12	0.18	0.37	0.48
CaO	2.66	4.02	0.28	0.19	0.60	0.12	1.45	0.36	1.36	0.99	0.90	1.19	0.17	0.43
MgO	0.83	0.78	0.06	0.03	0.07	0.03	0.04	0.04	0.09	0.09	0.09	0.15	0.03	0.19
Na_2O	3.72	3.82	2.91	4.23	3.11	4.46	2.77	3.97	3.81	3.49	3.66	4.70	3.34	3.13
K_2O	4.42	3.96	4.72	4.56	5.30	4.37	5.17	5.03	4.91	4.23	6.63	6.41	4.71	4.68
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00

NORMS

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Q	17.94	18.78	37.92	34.98	37.56	34.38	32.16	32.70	26.64	38.10	21.84	15.66	40.26	40.62
C	1.73	.....	3.47	.....	0.71	0.51	0.82	0.71	0.82	.....	0.82	.....	0.82	1.02
or	26.13	21.68	26.69	27.24	30.02	25.58	28.36	29.47	27.80	25.02	38.92	37.25	27.80	27.24
ab	31.44	31.96	23.58	35.63	25.15	37.20	22.01	33.01	30.92	29.34	30.39	39.30	27.77	26.30
an	12.79	15.57	1.39	1.11	2.50	0.56	6.39	1.95	6.39	4.73	4.45	3.34	1.11	1.95
Sal	90.03	87.99	93.05	98.96	95.94	98.23	89.74	97.84	92.57	97.19	96.42	95.55	97.76	97.03
di	.....	3.24	.....	.....	.....	.....	.....	.....	.....	.....	.....	0.65	.....	.....
wo	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	0.70	.....	.....
hy	2.00	0.40	0.10	.....	0.10	0.50	0.13	0.10	0.60	0.20	0.20	.....	.....	0.40
mt	.....	2.78	1.39	0.70	0.70	0.70	.....	0.70	1.86	0.46	.....	.....	1.39	1.62
hm	4.80	1.76	1.12	0.32	0.64	.....	2.24	0.80	.....	0.96	2.24	1.60	.....	.....
il	1.82	2.13	0.61	.....	0.15	.....	0.61	.....	0.46	0.46	0.15	0.61	.....	0.15
ru	0.24	.....	.....	.....	.....	.....	.....	.....	.....	.....	0.16	.....	.....	.....
Fem	8.86	10.31	3.22	1.02	1.59	1.20	2.98	1.60	2.92	2.08	2.75	3.56	1.39	2.17

No.

1. Quartz latite porphyry. Basal flow. Southwest side of Specimen Mountain.
2. Quartz latite porphyry. Boulder from base of agglomerate. Southwest side of Specimen Mountain.
3. Mud flow. Road cut near east end of Poudre Lake.
4. Rhyolite from boulder in mud flow on slope southeast of Poudre Lake.
5. Mud flow near "Saddle". Southwest side of Specimen Mountain.
6. Rhyolite in boulder from mud flow near "Saddle".
7. Breccia below lower pitchstone layer. Cliff near "Saddle".
8. Rhyolite from flow below lowest fault. Cliff west side Specimen Mountain.
9. Pitchstone. Lower layer, cliff near "Saddle".
10. Rhyolite. Top of cliff near "Saddle".
11. Rhyolite from flow. Cliff west of Iceberg Lake.
12. Rhyolite from plug. Peak half a mile north of top of Specimen Mountain.
13. Rhyolite from plug. Top of cliff west side of Specimen Mountain.
14. Rhyolite from plug. North side of Specimen Mountain.

quartz latite porphyry. It conforms in many respects to rocks classified as rhyobasalt by Johannsen (1932). Analyses are given in Table 1.

The presence of calcic plagioclase phenocrysts in a chemically intermediate rock suggests that the phenocrysts crystallized at depth. Partial resorption of the phenocrysts is evident and probably occurred during and after eruption.

PYROCLASTICS

Thick beds of tuff, ash, agglomerate, and mud flows record intermittent explosions of great magnitude. The materials in the pyroclastics reflect the composition of the laval in the plug at the time of explosion.

The oldest exposed explosive material is a thick bed of quartz latite ash resting directly upon the pre-Cambrian rocks and overlain by the basal flows of quartz latite porphyry. The ash consists of angular fragments of brown glass containing rare phenocrysts of labradorite. Similar ash beds alternate with the quartz latite flows.

Sandwiched between the older basal flows and younger rhyolite flows is a thick series of ash beds, agglomerates, and breccias. This series is well exposed in the cliff just west of the "Saddle". A rhyolite ash bed overlying the basal flows grades upward into an agglomerate bed about 200 feet thick. The agglomerate (Pl. 2, fig. 1) contains large rounded boulders of quartz latite and pre-Cambrian gneiss and granite embedded in rhyolite ash. Some of the boulders approach 3 feet in diameter and probably were ripped from the walls of the conduit at the time of eruption of the rhyolitic material in the matrix.

Overlying the agglomerate a bed of rhyolite ash about 100 feet thick grades upward into breccia. Tuff beds appear at several horizons in the section but rarely exceed 10 feet in thickness.

Plastic movement of ash and agglomerate as mud flows occurred at many stages in the history of the Specimen Mountain volcano. Water and steam derived perhaps in part from volcanic gases and in part from the atmosphere lubricated the movement and rapid flow of the unconsolidated pyroclastics down the sides of the volcanic cone. Solidified mud flows cover the south side of Poudre valley and stand out in exposures near the "Saddle".

Surface alteration of the pyroclastics is pronounced and has produced a variety of brilliant colors including green, red, orange, brown, and lavender. Alteration, probably attributable to the action of hot volcanic gases and water, has converted much of the rhyolitic ash into greasy hydrated bentonite. In wet seasons the bentonite is locally squeezed out on the surface in mud springs which dry into domical masses several feet high.

UPPER FLOWS

Flanking the plug on the south and west sides and truncated by it is a thick series of rhyolite flows and pitchstones alternating with beds of breccia and ash. The rhyolite is light gray to pinkish gray and generally shows conspicuous flow banding. In many exposures porous entirely crystalline layers alternate with dense or partly glassy layers.

Topaz is locally abundant in gas cavities in the porous rhyolite.

Microscopically the rhyolite is seen to contain abundant sanidine and quartz. The texture ranges from even-grained to porphyritic.

In the cliff west of the "Saddle" two prominent flows of black pitchstone crop out (Pl. 2, fig. 2). The lower one attains a maximum thickness of about 20 feet, the upper about 40 feet. The flows are separated by a layer of breccia which grades downward into the vesiculated reddish-brown top of the lower flow. A third less conspicuous layer of pitchstone is exposed near the top of the cliff. This flow grades laterally into a dense cryptocrystalline rhyolite. Locally the pitchstone is replaced by jasper nodules which have already been described (Wahlstrom, 1941). The pitchstone is shattered and highly fractured. Perlitic structures and lithophysae are locally present.

Under the microscope the pitchstone consists of brown glass with a wavy fluxion structure surrounding brecciated fragments containing an abundance of random to subparallel rodlike microlites. The glass has an average refractive index of about 1.55. Flow layering is emphasized by variations in the amount and state of oxidation of iron oxide in adjacent layers. An analysis of the lowest pitchstone layer is given in Table 1.

The pitchstone flows have the same chemical composition as the rhyolite flows and the plug rock except for a higher water content. The pitchstone is the rapidly chilled equivalent of the rhyolite flows and plug rock.

PLUG ROCKS

The plug marking the location of the conduit along which the lava moved toward the surface is roughly circular and consists chiefly of gray and brownish-gray rhyolite porphyry.

Field exposures show phenocrysts of translucent sanidine and smoky and clear quartz embedded in a glassy or fine-grained groundmass.

Locally at the contacts with older rocks, the plug rock grades into vitrophyre. Presumably the vitrophyre formed by rapid chilling of the lava as it moved through the conduit. The vitrophyre is jet black and contains conspicuous phenocrysts of sanidine and quartz.

Both the sanidine and the quartz are subhedral and, as seen under the microscope, rounded and embayed by resorption. The groundmass is commonly a brown glass showing swirling flow structures. In some of the plug rocks the groundmass is holocrystalline and contains lenticular aggregates of interlocking quartz and sanidine.

Tridymite is abundant in porous lenses and appears in clusters of tiny rectangular crystals. Probably the finer portions of the groundmass contain considerable proportions of this mineral.

Euhedral oligoclase makes up about 5 per cent of the rock. In addition a few small flakes of strongly pleochroic biotite and grains of magnetite are present. Hornblende is rare and generally shows strong resorption which has resulted in the conversion of the hornblende to magnetite and pale biotite.

Several analyses are given in Table 1 and support the classification of the plug material as rhyolite porphyry.

DIFFERENTIATION

Despite the diversity of textures of the flow and plug rocks and the variations in the state of subdivision of the pyroclastics, the volcanics of Specimen Mountain chemically and mineralogically comprise only two types. Older quartz latitic rocks probably are parental to the younger rhyolitic effusive and plug rocks.

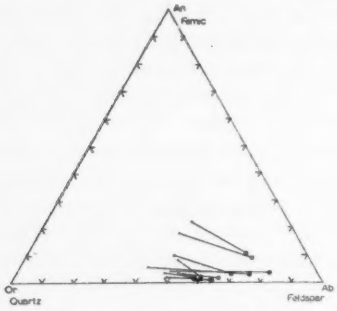

FIGURE 6. *Larsen plots of Specimen Mountain rocks*

Dots are normative or-ab-an; circles are normative quartz-feldspar-femic.

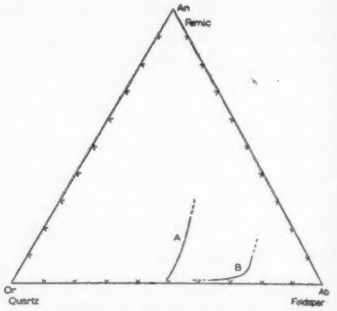

FIGURE 7. *Triangular diagram*

Showing generalized differentiation trends of Specimen Mountain rocks.

A. Normative feldspars.

B. Normative quartz-feldspar-femic.

Spock (1928) reports basaltic and andesitic rocks in the near-by Never Summer Range, and these rocks may be part of a differentiation series which culminated in the eruption of quartz latite and rhyolite from the Specimen Mountain volcano.

Figure 6 shows a ternary Larsen plot of the chemical analyses of the Specimen Mountain rocks. The positions and attitudes of the plots accord fairly well with the silicic rocks of the San Juan Mountains in southwestern Colorado (Larsen, 1938).

The plot in Figure 6 is based on a modification of the norms of the analyzed rocks. First the normative feldspars are recalculated to 100 per cent, and the points are plotted; then the normative quartz, feldspar, and femic minerals are calculated to 100 per cent, plotted, and lines are drawn to join the two points obtained for each analysis.

Figure 7 shows the general trends of crystallization for the feldspars and quartz-feldspar-femic minerals. The diagram suggests that the rhyolite was derived from quartz latite by normal gravitative differentiation at depth. Crystallization and subtraction of crystals of calcic plagioclase and ferromagnesian minerals from a quartz latitic magma would lead to the formation of a rhyolitic magma exactly like the younger effusives in Specimen Mountain.

This conclusion is supported by thin sections of the two rock types. The quartz latite contains abundant embayed phenocrysts of calcic plagioclase and euhedrons of hornblende and pyroxene. The latter minerals have been converted in part to biotite and magnetite. The reaction of the magmatic liquid with both the feldspar and the ferromagnesian minerals suggests that these minerals crystallized from the magma at depth.

The rhyolite contains only a few flakes of biotite and about 5 per cent oligoclase.

PNEUMATOLYTIC AND HYDROTHERMAL ACTIVITY

The more porous layers in the rhyolite flows contain abundant tridymite and, locally, clusters of crystallized topaz. Both minerals probably formed almost immediately after the extrusion of the enclosing rocks and were deposited from mineralizer-rich concentrations of liquids and gases resulting from the rapid crystallization of anhydrous silicates from a water-rich magma.

Later than the tridymite and topaz and localized by faults and joints are deposits of cryptocrystalline silica. The silica-bearing solutions probably came from the deeper, still hot portions of the volcano and moved toward the surface along the faults.

Locally the fault zones are silicified by red or gray chalcedony. In the obsidian layers where they are cut by faults, silicification has produced numerous geodes partially or completely filled with onyx and opal.

AGE

Sufficient data are not at hand to permit the determination of the exact geologic age of the Specimen Mountain rocks. However, it is safe to assume that they represent one phase of the complex Cenozoic igneous history of the Front Range of Colorado.

Exposed tuff beds examined in an effort to unearth identifiable fossils yielded only poorly preserved carbonized plant impressions. Further search may lead to the discovery of a fossil flora or fauna sufficient to permit an age assignment.

Most of the Cenozoic sediments in the intermontane basins and in the plains area east of the Front Range contain abundant volcanic materials. Further information regarding the composition and distribution of these materials may be necessary before the Specimen Mountain eruptions can be dated.

The freshness of the exposures and the rapid rate at which the extruded rocks are now being eroded seem to militate against an early Cenozoic age. Glacial erosion of the plug rocks and flanking effusives indicates that the volcanic rocks are pre-Pleistocene. Tentatively, then, a Miocene or Pliocene age appears plausible.

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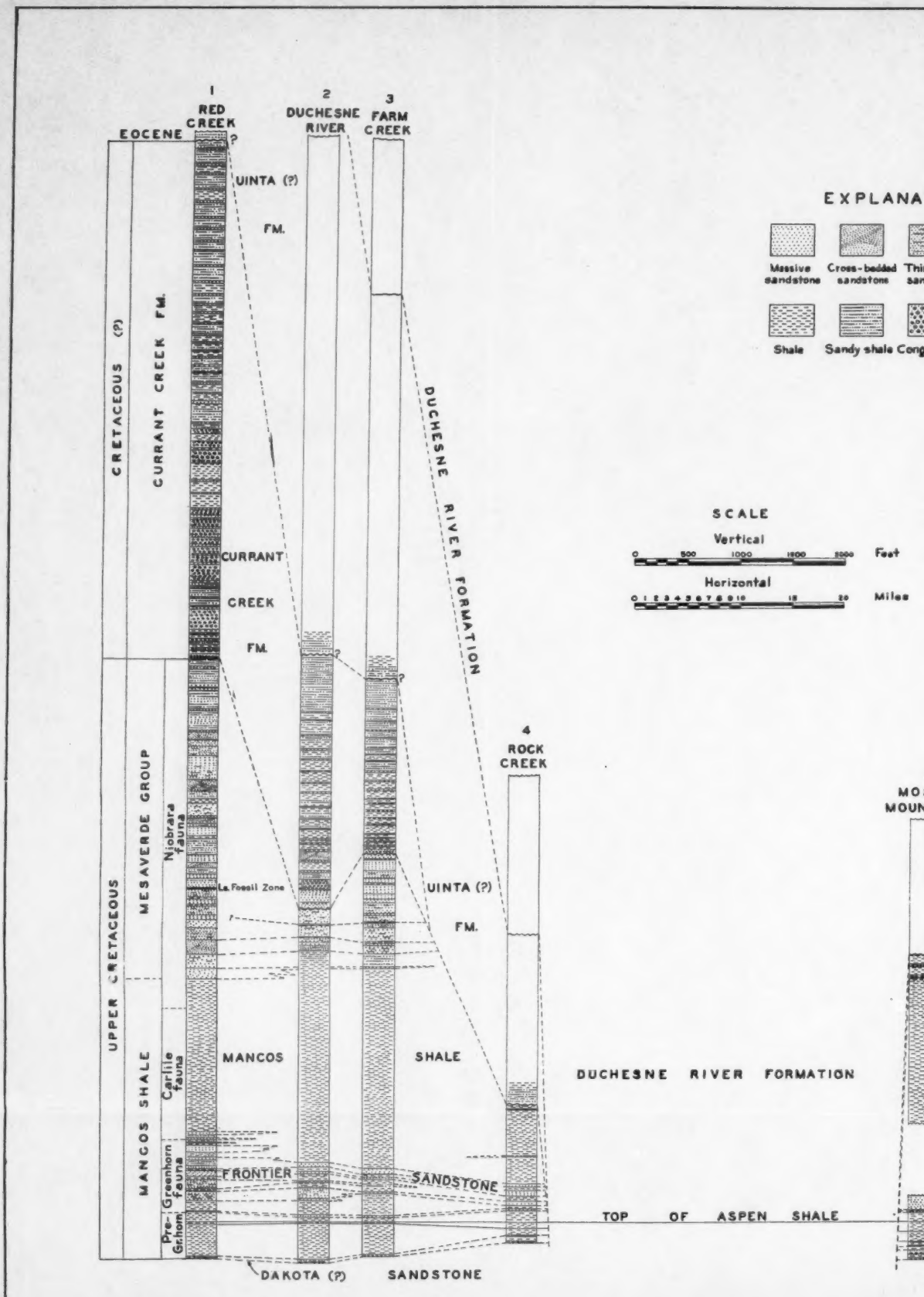
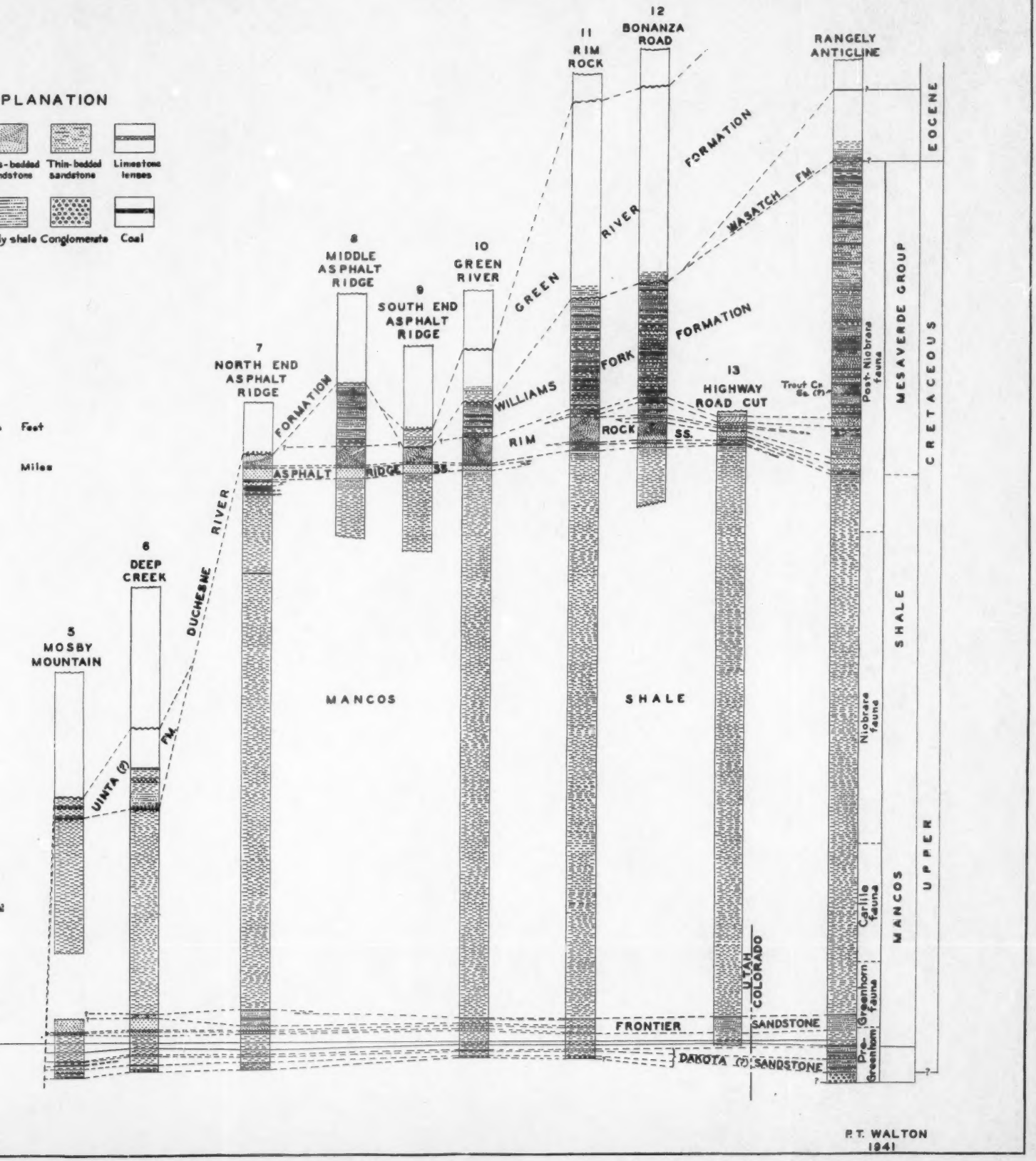


UPPER CRETACEOUS STRAT



US STRATIGRAPHY OF THE UINTA BASIN, UTAH

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GEOLOGY OF THE CRETACEOUS OF THE UINTA BASIN, UTAH

BY PAUL T. WALTON

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ABSTRACT

The Cretaceous rocks of the Uinta Basin, Utah, are exposed along the south flank of the Uinta Mountain anticline and are overlain unconformably by Tertiary sediments throughout this area.

The Dakota (?) sandstone of this paper disconformably overlies the Jurassic Morrison and is overlain conformably by the Mancos shale. It probably represents rocks of both Upper and Lower Cretaceous age.

The Mancos shale, as mapped and studied, comprises only a lithologic unit since in the Vernal vicinity it is equivalent in age and stratigraphic position to the Mesaverde group sandstones of the Tabby Mountain region to the west. The Mancos is divided into the following members, all easily identified throughout the Basin: lower shale member, Aspen shale, middle shale member, Frontier sandstone, and upper shale member.

The Mesaverde group refers only to a lithologic facies since at the west end of the area it is predominantly Niobrara in age, whereas at the east it is completely post-Niobrara. In the Vernal region it is divided into three members: the Asphalt Ridge sandstone, the Rim Rock sandstone, and the Williams Fork formation (contains more than type Williams Fork). The lower members of the Mesaverde group intertongue with the upper portion of the Mancos shale and wedge out toward the east.

The name Currant Creek is proposed for a sequence of conglomerate beds unconformably overlying the Mesaverde group in the Tabby Mountain region. This new formation is tentatively referred to the Upper Cretaceous but it probably represents the transitional strata deposited from late Upper Cretaceous to Eocene during the early phases of the Laramide folding.

The Uinta Basin is an asymmetrical syncline with its steep side along the flank of the Uinta Mountain anticline to the north from which numerous small noses and supplementary folds plunge southward. The Basin's structural history involves the following movements: local uplift in the west end in early Montana time; folding, progressively from west to east, during late Cretaceous to early Eocene; first uplift during post-Green River (Eocene)—pre-Uinta (Eocene); second uplift during post-Uinta (Eocene)—pre-Duchesne River (Oligocene); regional tilting in post-Duchesne River (Oligocene)—pre-Bishop (Miocene?); subsidence in post Bishop (Miocene?)—pre-Quaternary; continental uplift in late Tertiary or early Quaternary.

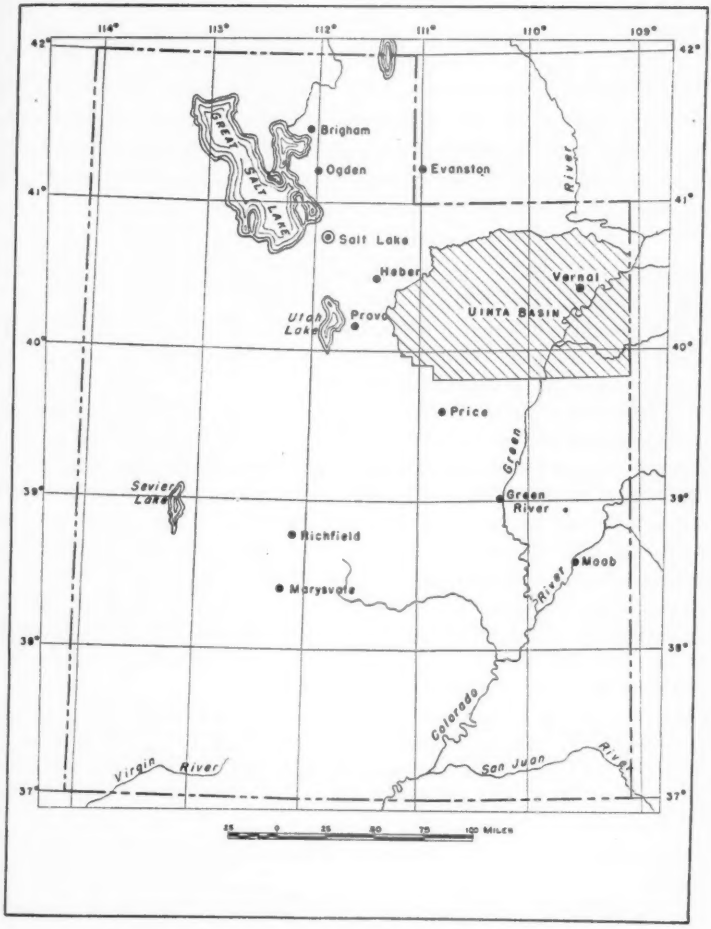


FIGURE 1.—*Index map showing location of Uinta Basin, Utah*

INTRODUCTION

GENERAL FEATURES OF THE AREA

The Uinta Basin of Utah is a geographic and geologic province in northeastern Utah roughly 6000 square miles in area (Fig. 1). Its natural geographic boundaries are the Uinta Mountains on the north and the Roan Cliffs on the south, rising to a maximum relief of 8000 and 5000 feet, respectively, above the floor of the Basin.

The Basin is a structural and topographic low. Although it did not assume its present form until post-Oligocene time, it was part of a much larger intermontane basin which came into existence near the close of the Cretaceous, during the first

phases of the Laramide Revolution. As a consequence, the Tertiary sediments overlap the beveled edges of all formations from the Algonkian to the Cretaceous. The continuity of many of the Mesozoic formations along the south flank of the Uinta Mountains is broken because of this overlap, so that a study of the details of their stratigraphy is difficult. This is probably one reason why no comprehensive studies on the Cretaceous section as a whole have ever been made and also why no detailed correlation between this area and the surrounding regions has ever been worked out.

PREVIOUS GEOLOGIC WORK

The earliest published account of explorations in the Uinta Basin area of Utah is found in Fremont's travels (1842–1844, p. 277–280).

The first general description of the geology of the area was given by Powell (1876, p. 1–218) who described "Uinta structure" as a flat-topped plateau-like uplift bounded by monoclinal flexures on either side.

The geologists of the 40th Parallel Survey (King *et al.*, 1877) mapped the geology in a reconnaissance manner in 1869 and 1871. In 1903 Berkey (1905, p. 516–530) mapped a portion of the Duchesne River valley showing the Cretaceous rocks and coal outcrops. Weeks (1906, p. 427–448) encircled the Uinta Mountains in 1906 and made a revised geologic map on which he further differentiated the various previously described formations. In 1906 Gale (1906, p. 1–265) completed his investigations along the northeastern flank of the Uinta Basin as far west as Little Mountain, connecting with the Danforth Hills and Grand Hogback coal fields.

The Cretaceous coal-bearing strata were studied in some detail by Lupton (1912, p. 579–594, 592–628) in 1912. He mapped and traced out the important coal beds of the Tabby Mountain and Deep Creek coal fields.

More recently, the bituminous sandstones of Asphalt Ridge were studied by Spieker (1930, p. 77–98). A comprehensive study of the Tertiary formations of the Utah part of the Uinta Basin was made by Kay (1934, p. 357–372), and more recently Forrester (1937, p. 631–666) completed his work on the structure of the Uinta Mountains.

Tolmachoff (1942, p. 41–61) described Cretaceous fossils found in Asphalt Ridge near Vernal.

All other geological investigations in adjoining regions bearing upon features of this report are cited at the proper place.

PRESENT INVESTIGATION

The field work for this report was done during the period of June to November in 1941. The original purpose of the investigation was to determine the oil possibilities of the Cretaceous strata of the area. This necessitated a detailed study of the stratigraphy and structure in addition to the investigation of possible source and reservoir rocks. The economic geology of these strata, however, will not be discussed here. A second object of this investigation was to correlate the Cretaceous of the Uinta Basin with that of the Book Cliffs-Wasatch Plateau area where much detailed work has been done by Clark, Spieker, Fisher, and Erdmann.

The Forest Service *Base map of the Uinta Basin, Utah* was used for final plotting

of the results of the field work. Field mapping was done on the available U. S. Geological Survey topographic maps and General Land Office plats by triangulation and traverse with a Brunton compass from land survey section corners. Stratigraphic measurements were made by triangulation, pacing, and differential leveling.

ACKNOWLEDGMENTS

The writer is very grateful to Professors Robert R. Shrock and Walter L. Whitehead of the Massachusetts Institute of Technology for supervision of the study. Their many suggestions concerning the problem and their invaluable assistance in preparing the report are greatly appreciated.

Professor Edmund M. Spieker of The Ohio State University was most generous in giving of his great store of information concerning the geology of this area and the surrounding region. This aid was of vital assistance in correlating the results of this study with the regional geology. His thorough criticism of the manuscript was very helpful in the preparation of the final report.

Dr. John B. Reeside, Jr., of the U. S. Geological Survey identified the fossils and generously discussed by correspondence the problems concerning the age and correlation of the Cretaceous stratigraphy of the region.

Dr. I. P. Tolmachoff of the Carnegie Museum kindly supplied advance information from his paper then in press; and Mr. J. LeRoy Kay, of the same institution, gave much information concerning the Tertiary formations of the area.

Dr. A. A. Baker of the U. S. Geological Survey gave helpful suggestions concerning the geology of the Strawberry Valley region, and Dr. W. H. Bradley, of the same organization, generously discussed by correspondence the relationships of the Tertiary formations of the region.

STRATIGRAPHY

The geologic formations of the Uinta Basin range in age from Algonkian to Quaternary. The total thickness of the sequence is roughly 40,000 feet in the eastern part and 48,000 feet in the western part of the Basin. The exposed strata include quartzite, sandstone, limestone, shale, and conglomerate and were deposited in environments ranging from shallow marine to terrestrial.

The stratigraphy of the region has been discussed in numerous publications. This report will consider only the Cretaceous and immediately contiguous rocks. It seems appropriate, however, to review briefly the events preceding their deposition.

The Uinta Mountains geosyncline probably came into existence in the Proterozoic or earlier, and about 12,000 feet of Algonkian arenaceous sediments were deposited without interruption. As no regional unconformity separates these beds from the succeeding Cambrian formations there does not seem to have been orogenic movement at the close of the Proterozoic era. Since, however, the Cambrian is overlain disconformably by Mississippian limestone, uplift seems indicated during the Cambro-Mississippian interval. Following the Mississippian limestone conformably are the Pennsylvanian quartzites and intercalated series and the Permian formations. No unconformity has ever been reported between the Triassic and the Permian strata, implying that another era, the Paleozoic, closed without orogenic movement

in this area. The Triassic red beds are followed conformably by the Jurassic continental, marine, and fluviatile strata on which the Upper Cretaceous Dakota (?) sandstone lies disconformably. An apparently continuous succession of Cretaceous rocks overlies the Dakota (?) in the eastern part of the Basin. A somewhat similar and probably roughly contemporaneous sequence in the Tabby Mountain area to the west is interrupted by an unconformity which separates the underlying beds of Niobrara age from an overlying unfossiliferous succession of conglomerates, sandstones, and shales of probable Cretaceous age.

JURASSIC SYSTEM

The Morrison formation outcrops along the south flank of the Uinta Mountains and, being composed of multi-colored rocks, its top affords a well-marked boundary for the base of the Upper Cretaceous. Characteristically, it is composed of pink, red, gray, purple, maroon, yellow, and green shales with occasional interbeds of coarse brown and buff sandstone. This formation contains the dinosaurs at Dinosaur National Monument on the Green River above Jensen. It maintains, as a whole, a constant thickness of about 780–800 feet throughout the area.

A fluviatile origin seems probable for most of this formation, although some of the individual beds seem to be marine.

The Morrison produces commercial quantities of gas in the Ashley Valley gas field.

UPPER CRETACEOUS SYSTEM

DAKOTA (?) SANDSTONE

GENERAL FEATURES: The strata here described as the Dakota (?) sandstone are considered as the base of the Upper Cretaceous section in this region. They lie immediately above the variegated shale of the Morrison and below the lower shale member of the Mancos. This sandstone outcrops as a cuesta in several places along the south flank of the Uinta Mountains.

LITHOLOGIC CHARACTER AND THICKNESS: In the western part of the Uinta Basin the Dakota (?) sandstone is a coarse-grained, partly cross-bedded, buff sandstone about 30–50 feet thick, lying concordantly on the Morrison, so that it appears as the uppermost member of that formation. Farther east it coarsens and thickens and splits into several beds of coarse sandstone separated by sandy shale (Pl. 1). Near the Colorado border, the sandstone beds coalesce, and the section shows the following lithology:

Section of Dakota (?) sandstone at Sand Springs, Colorado

	Feet
Lower shale member of the Mancos shale: shale, dark gray.	
Dakota (?) sandstone	
Sandstone, massive, buff, interbedded with gray sandy shale	78
Conglomerate and coarse sandstone	20
Shale, sandy with a few thin beds of brown, fine-grained sandstone	127
Conglomerate, coarse pebbles	103
Total	328
Morrison formation: variegated shale and brown sandstone.	

The conglomeratic beds of the above section are composed largely of pebbles and sand grains of brown quartzite and black chert and are present mainly in the

TABLE 1.—*Stratigraphy of the western part of the Uinta Basin, Utah*

Formations concerned in this paper

Age	Group and formation	Member	Character of Beds	Thickness (feet)
Miocene(?)	Bishop conglomerate Unconformity		Conglomerate, boulders 1 to 6 feet in diameter, intermixed with gravel and sand.	0–500
Oligocene	Duchesne River formation Unconformity		Sandstone, buff, gray; shale, red, pink, sandy; conglomerate. Fluviatile.	1500
Eocene	Uinta and Uinta (?) formation Unconformity		Sandstone, light brown, gray; shale, gray, green, reddish brown, red, pink, sandy. Fluviatile.	1800–5400
	Green River Shale Unconformity?		Shale, gray, greenish gray, oil shale in middle of section, paper shales; sandstone, buff. Lacustrine.	0–5000
Cretaceous(?)	Currant Creek formation Unconformity		Conglomerate, pebbles up to 14 inches in diameter; sandstone, yellow, buff, massive; shale, pink, white, sandy. Fluviatile.	0–4822
Upper Cretaceous.	Mesaverde group		Sandstone, buff, white, massive; shale, a few thin beds, gray, sandy. Marine at base. Continental, coal-bearing at top. Contains Niobrara fossils throughout most of the section.	0–3001
	Mancos shale	Upper shale member	Shale, gray, drab, clay. Not well exposed. Marine.	800–2000
		Frontier sandstone	Sandstone, buff, brown, white; shale, gray, black, sandy; coal; brackish water and marine. Contains Greenhorn and Carlile fossils.	248–756
		Middle shale member	Shale, yellowish, drab, sandy. Marine.	42–108
		Aspen shale	Shale, siliceous, dark gray, splintery. Marine. Contains fish scales. Weathers silver gray.	15–60

TABLE 1.—*Continued*

Age	Group and formation	Member	Character of Beds	Thickness (feet)
Upper Cretaceous.—*Continued*	Mancos shale—*Continued*	Lower shale member	Shale, drab, locally dark gray. Marine.	74–310
	Dakota(?) sandstone Unconformity		Sandstone, brownish, gray, buff, conglomeratic; shale, gray, sandy. Marine at top.	30–48
Jurassic.	Morrison formation		Shale, pink, purple, gray, yellow, maroon, green; sandstone, brown, buff, coarse. Mostly fluviatile.	780–800

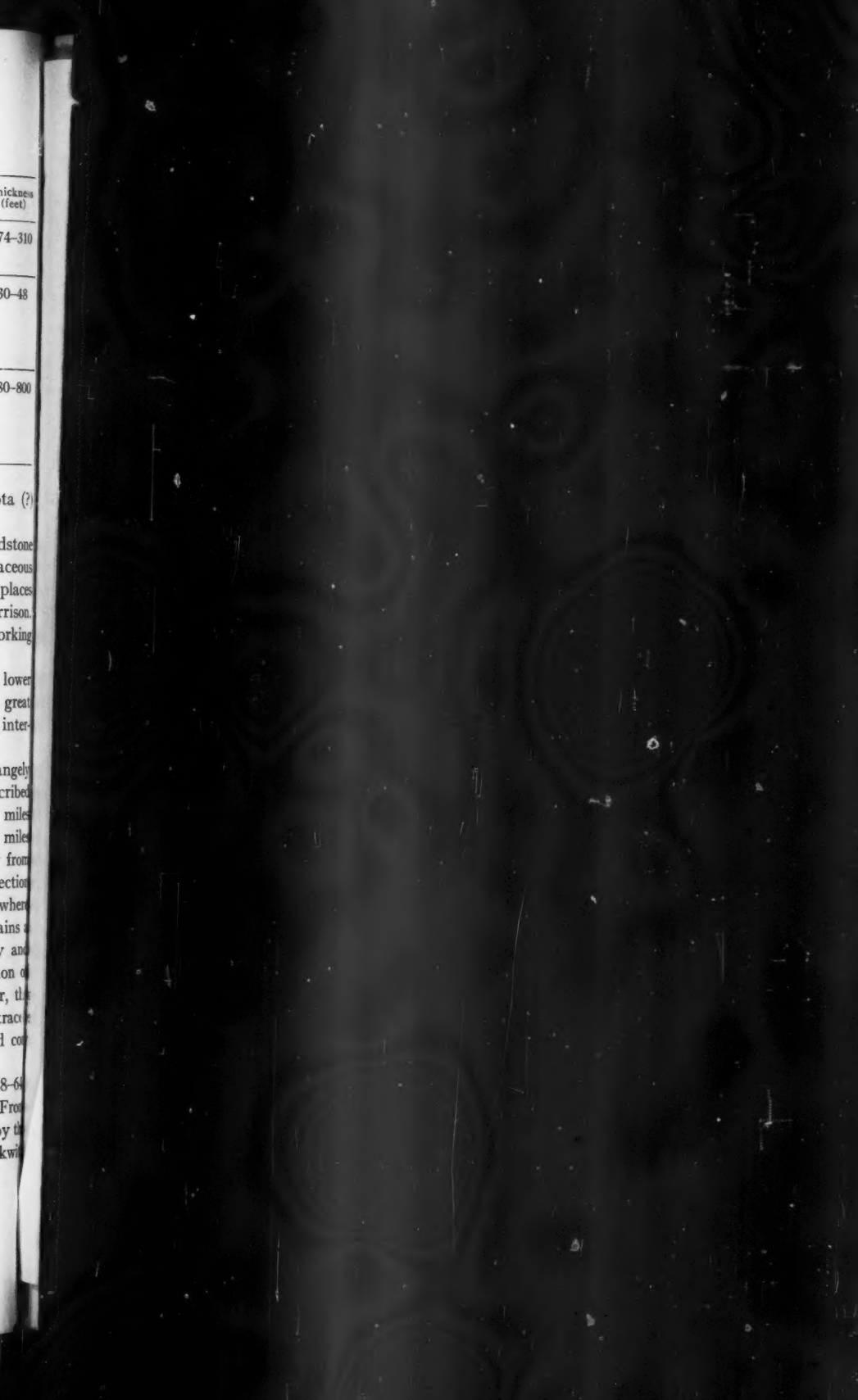
easternmost part of the Basin. Elsewhere throughout the area, the Dakota (?) is a coarse sandstone.

STRATIGRAPHIC RELATIONS: The lower boundary of the Dakota (?) sandstone appears to represent the base of a transgressive overlap formed when the Cretaceous sea advanced westward over the underlying Morrison formation. In many places the Dakota (?) probably represents little more than the reworked top of the Morrison. The absence of marine fossils in the Dakota (?) suggests that most of this reworking was done by streams.

The upper part of the Dakota (?) sandstone is succeeded conformably by the lower shale member of the Mancos. This shale represents the initial deposit of a great thickness of marine Upper Cretaceous sediment which was laid down without interruption during a long period of steady downsinking.

AGE AND CORRELATION: The Dakota (?) sandstone in the vicinity of the Rangely anticline is very similar lithologically and stratigraphically with beds described as Dakota (?) by Erdmann (1934, p. 27–28) in the Book Cliffs coal field 80 miles to the south and by Sears (1924, p. 285–286) in Moffat County, Colorado, 20 miles to the north. The Dakota (?) sandstone can be traced almost continuously from the Wasatch Plateau eastward throughout the Book Cliffs to Erdmann's section (Pl. 3). The Dakota (?) of this paper can be traced continuously, except where overlain by Tertiary formations, along the south flank of the Uinta Mountains a similar distance (Pl. 1), and throughout it is nearly identical lithologically and stratigraphically with the Dakota (?) of the Book Cliffs. At least a portion of each of these three sandstone formations are definite correlatives. However, the thick basal conglomerate bed shown in the Sand Springs section cannot be traced far in any direction. It is suggested that it may be Lower Cretaceous and correlative to the Cloverly group of Wyoming.

North of the Uinta Mountains in southwestern Wyoming, Veatch (1907, p. 58–6 found beds referable to the Dakota (?) sandstone, as it is known west of the Fro Range in Colorado. In that area, the basal Upper Cretaceous is represented by th Bear River formation, which overlies the Jurassic and Lower Cretaceous Beckwi



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Strawberry River
Green River
Red Creek
Fruitland
Strawberry Reservoir
WASATCH CO.
DUCHESNE CO.
UTAH CO.
R10W
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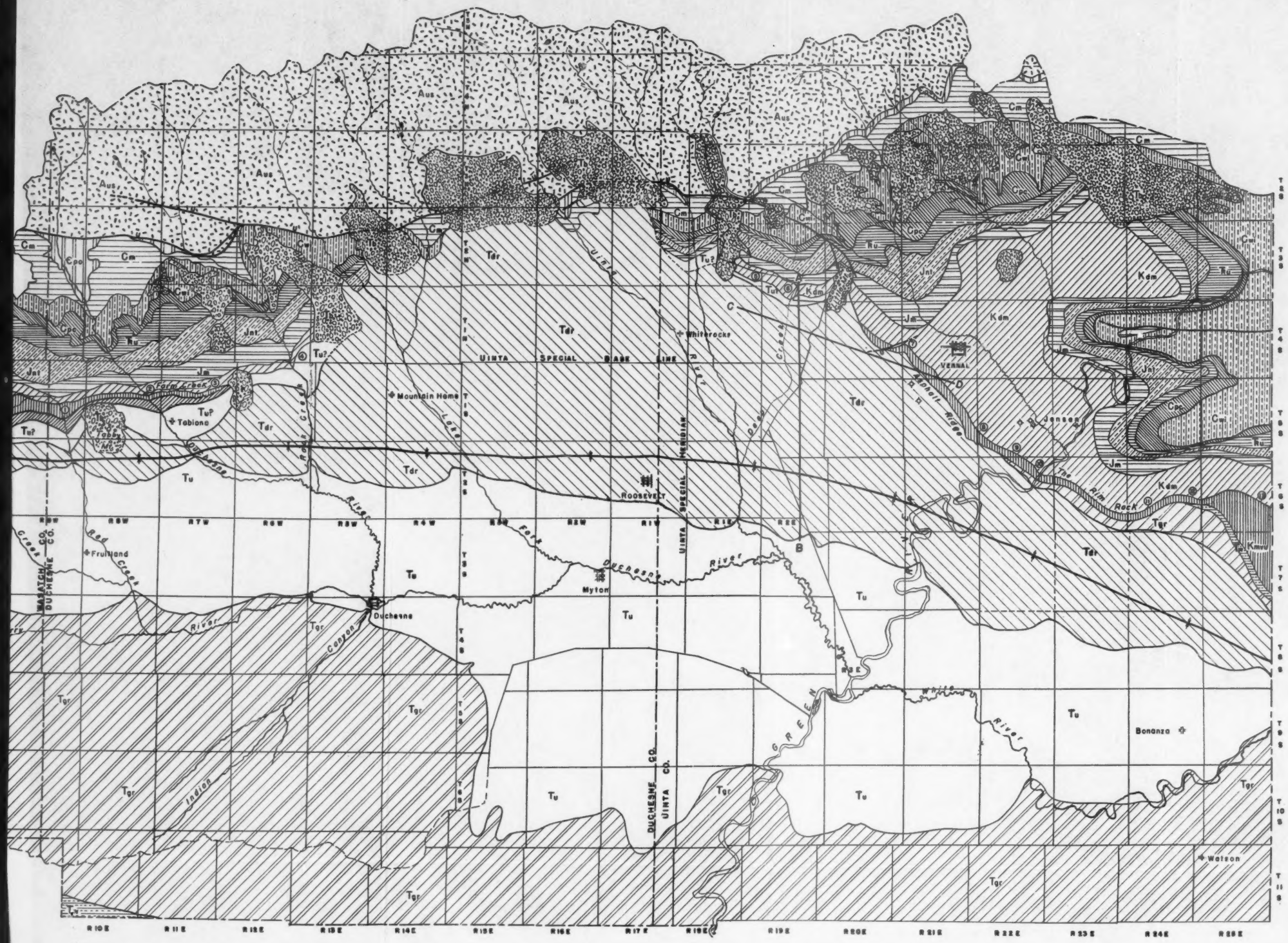
Section A-B



EXPLANATION

SEDIMENTARY ROCKS

TERTIARY

- OLIGOCENE MIOCENE(?)
  - Tb — Bishop
  - Tdr — Duchesne River
- EOCENE
  - Tu — Uinta
  - Tgr — Green River
  - Tw — Wasatch

UPPER CRETACEOUS

- Kmv — Current Creek
- Kmvu — Mesa Verde Undivided
- Kdm — Dakota Mancos

JURASSIC

- Jm — Morrison
- Jnt — Nugget Twin Creek

TRIASSIC

- Tu — Triassic Undivided

CARBONIFEROUS

- PERMIAN
  - Cpc — Park City
- PENNSYLVANIAN
  - Cwi — Weber & Intercalated Series
- MISSISSIPPIAN
  - Cm — Madison

CAMBRIAN

- Cpo — Pine Valley Ophir

ALGONKIAN

- Aus — Uinta Series

IGNEOUS ROCKS

TERTIARY

- Tan — Andesite Extrusives

① Location of sections Shown on Plate 1

◆ Wells penetrating the Cretaceous

Geology of the Jurassic to the Tertiary by P.T. Walton. Additional information compiled from the works of Forrester, Kay, Lupton, Spieker, Bradley, Sears & others.

Section A-B

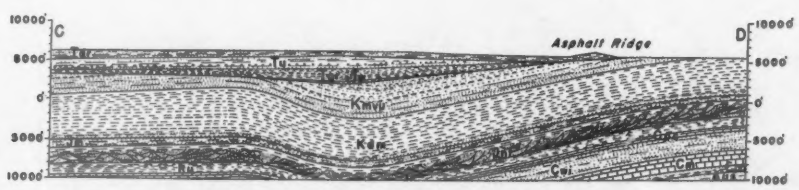
Scale

Map

0 1 2 3 4 5 6 7 8 9 10 Miles

Geologic Sections

0 5000 10000 15000 20000 Feet



Section C-D

GEOLOGIC MAP AND SECTIONS OF THE UINTA BASIN, UTAH

Miocen
Oligoc
Eocene
Upper
ceou

TABLE 2.—*Stratigraphy of the eastern part of the Uinta Basin, Utah*

Formations concerned in this paper

Age	Group and formation	Member	Character of Beds	Thickness (feet)
Miocene(?)	Bishop conglomerate. Unconformity		Conglomerate, boulders 1 to 6 feet in diameter, intermixed with gravel and sand.	0–500
Oligocene	Duchesne River formation Unconformity		Sandstone, buff, gray; shale, red, orange red; conglomerate. Fluviatile.	1372
Eocene	Uinta and Uinta (?) formation Unconformity		Sandstone, buff, gray; shale, gray, reddish brown, red, yellow; conglomerate. Fluviatile.	700–1648
	Green River shale Unconformity?		Shale, gray, greenish gray, oil shale in middle of section, paper shales; sandstone, buff, hard, blocky. Lacustrine.	1800–2400
	Wasatch formation Unconformity		Shale, gray, pink, green sandstone, thin-bedded, brown. Lacustrine.	780
	Mesaverde Group: Williams Fork formation		Sandstone, buff, brown, white; shale, gray, pink, brown; coal. Estuarine and brackish water. Includes more than type Williams Fork.	0–2614
	Mesaverde Group: Rim Rock sandstone		Sandstone, gray, speckled cross and thin-bedded; shale tongues. Marine.	112–416
	Mesaverde Group: Asphalt Ridge sandstone		Sandstone, yellowish, white, soft, massive. Marine. Contains post-Niobrara fossils.	0–101
Upper Cretaceous.	Mancos shale	Upper shale member	Shale, drab, yellowish, blue, clay; sandstone, few brown thin beds at top. Marine.	5000–6000
		Frontier sandstone member	Sandstone, buff, brown, white; shale, gray, black, sandy; coal. Marine and brackish water.	163–215
		Middle shale member	Shale, dark, carbonaceous, yellow, drab, variable. Marine.	79–98

TABLE 2.—*Continued*

Age	Group and formation	Member	Character of Beds	Thickness (feet)
Upper Cretaceous.—*Continued*	Mancos shale—*Continued*	Aspen shale	Shale, siliceous, dark gray, splintery. Marine. Contains fish scales. Weathers silver gray.	36–95
		Lower shale member	Shale, gray, locally dark, yellowish gray. Marine.	26–91
	Dakota (?) sandstone Unconformity		Sandstone, brown, buff, coarse, conglomeratic; shale, yellowish, gray, sandy; conglomerate. Base may be Lower Cretaceous.	169–328
Jurassic	Morrison formation		Shale, pink, purple, gray, yellow, maroon, green; sandstone, buff, brown, coarse. Mostly fluviatile.	780

formation. Neither the Bear River nor its typical fauna has ever been identified south of the type area although it has a maximum thickness of 5000 feet. Veatch suggested that it is an estuarine deposit that was separated by barriers from the transgressive sea, so that typical Dakota (?) beds were never laid down in that region. It appears, therefore, that the base of the Upper Cretaceous cannot be correlated from the north flank to the south flank of the Uinta Mountains.

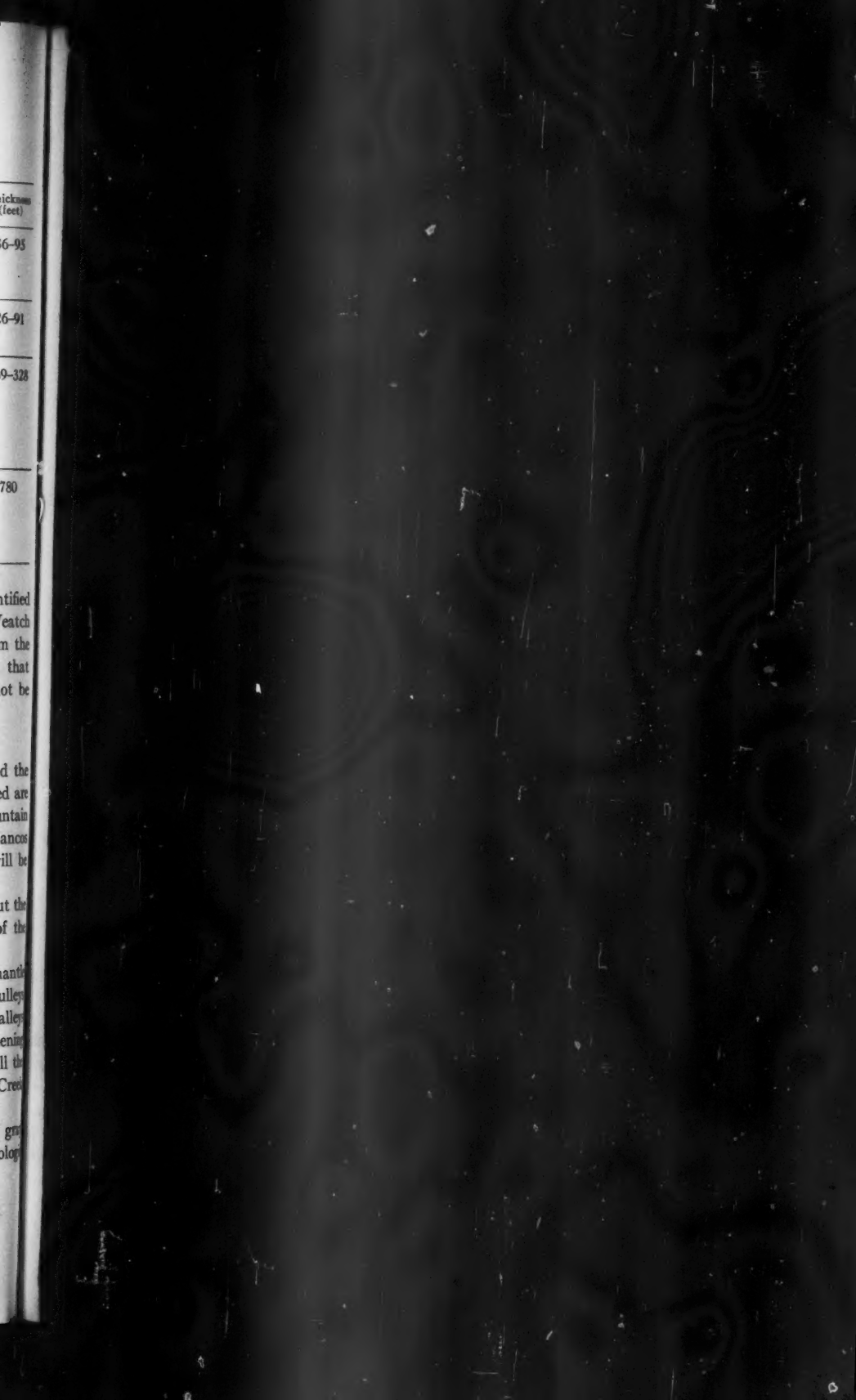
MANCOS SHALE

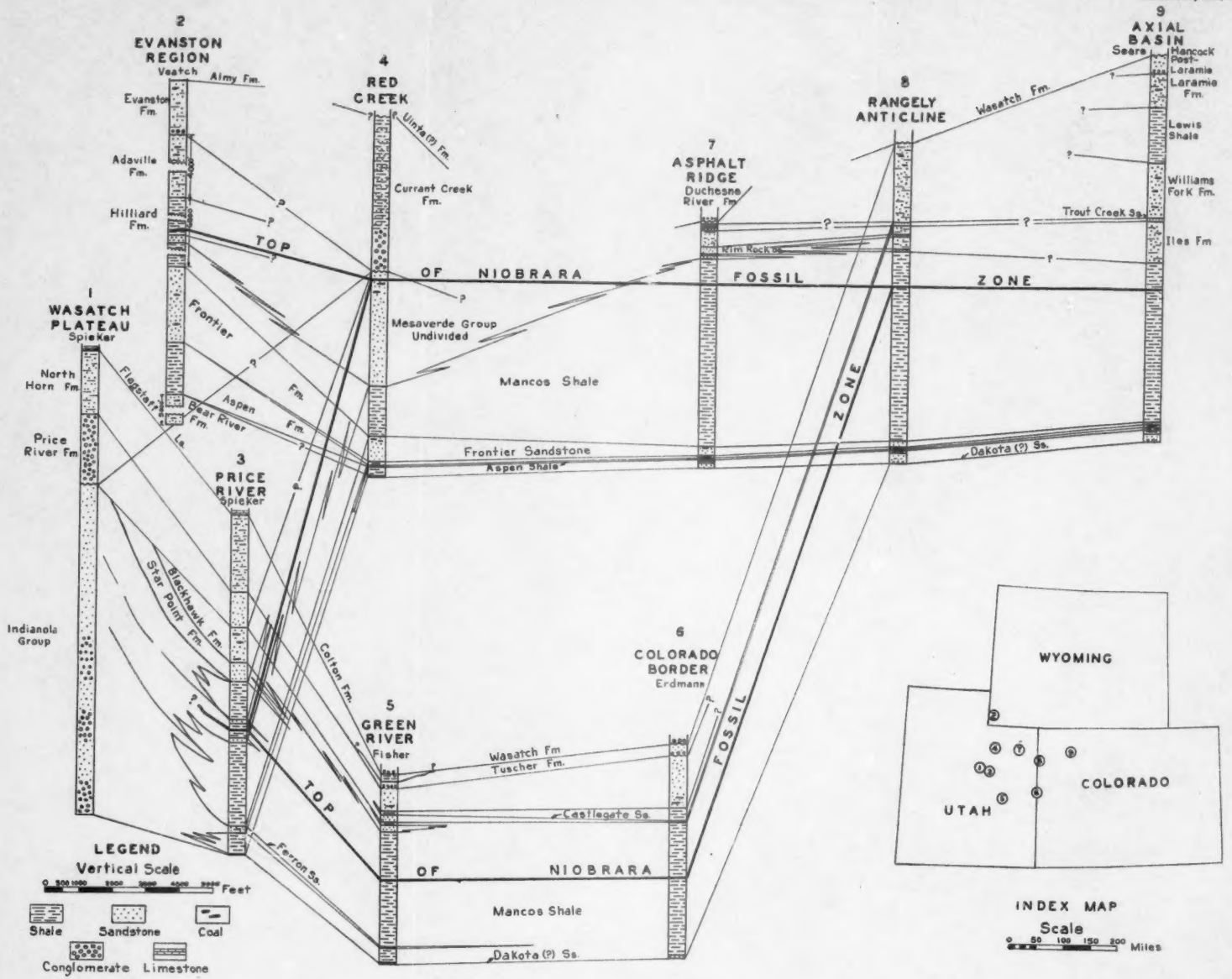
GENERAL FEATURES: All the strata between the Dakota (?) sandstone and the lowest member of the sequence of sandstone mapped as Mesaverde Undivided are designated as Mancos shale. Actually, sandstone strata in the Tabby Mountain region mapped as Mesaverde are the equivalent of the main body of the Mancos shale near Vernal (Pl. 1). Notwithstanding, only the argillaceous facies will be considered Mancos shale in this report.

The Mancos shale conformably overlies the Dakota (?) sandstone throughout the Uinta Basin but intertongues at its upper boundary with the sandstones of the Mesaverde group.

Outcrops of the shale are uncommon because it readily weathers into a mantle of clay soil. This soil, unless mixed with sand, is easily eroded, and large gulleys are common. Because of its great thickness, the Mancos shale forms broad valleys wherever the dip is sufficiently low. Such is the case at Vernal where the flattening of dip on a terracelike structure produces the wide Ashley Valley in which all the cultivated land occurs on outcrops of the shale. This is true also at Deep Creek and Farm Creek.

LITHOLOGIC CHARACTER AND THICKNESS: Most of the formation consists of gray to drab marine clay shale, but it contains several members of differing lithology





P. T. Walton 1942

REGIONAL CRETACEOUS CORRELATION DIAGRAM

character. The total thickness varies from west to east, being 800 feet at Currant Creek and more than 6000 feet near the Utah-Colorado border (Pl. 10, fig. 1).

General Section (1)[1] of the Mancos shale at Red Creek

Mesaverde group, undivided: buff, massive sandstone.

Mancos shale	Feet
Shale, drab, clay (Upper shale member)	1450
Sandstone, white, brown, massive (Frontier sandstone member)	756
Shale, dark, clay (Middle shale member)	100
Shale, siliceous, dark gray (Aspen shale member)	15
Shale, black, clay, (Lower shale member)	310
Total	2631

Dakota (?) sandstone: buff coarse sandstone.

The members are recognizable everywhere in the area, but there is considerable variation in their thickness from east to west as shown by the following section:

General Section (11) of the Mancos shale at the Rim Rock

Rim Rock sandstone: sandstone, gray, some cross-bedding.

Mancos shale	Feet
Shale, gray, drab, clay (Upper shale member)	5800
Sandstone, buff, white; shale, gray, sandy (Frontier sandstone member)	157
Shale, dark, clay (Middle shale member)	79
Shale, siliceous, dark gray, contains fish scales, splintery (Aspen shale member)	36
Shale, black, carbonaceous, clay (Lower shale member)	43
Total	6115

Dakota (?) sandstone; buff, conglomeratic.

MEMBERS OF THE MANCOS SHALE: The five members of the Mancos shale shown in Tables 1 and 2 and in the sections above are fairly constant in appearance and lithologic character wherever they outcrop.

The lower shale member lying directly above the Dakota (?) sandstone is distinctly different from the other members, particularly in the eastern part of the Uinta Basin near the Colorado border, where it is typically a black band 50 to 100 feet thick outcropping between the Dakota (?) and Frontier sandstones. Farther west along Red Creek, this shale is gray to drab, and it is 310 feet thick.

In its dark color, marked fissility, and identical stratigraphic position, this shale resembles greatly the black shale found between the Dakota (?) and Ferron sandstones in Castle Valley, described in several publications, probably best by Spieker (1931, p. 19). The writer has examined the shale at the latter locality, and the resemblance is striking. At least tentative correlation between the two seems possible.

Gale (1910, p. 61–62) has described strata of the same lithologic character and stratigraphic position in northwestern Colorado a few miles north of where the writer terminated his stratigraphic traverse. A definite correlation therefore seems possible between these two general areas.

Immediately above the lower shale member of the Mancos is a conspicuous dark-gray siliceous shale member, the Aspen, that was used as a datum plane for measure-

[1] Numbers refer to number of section (Pl. 1) and location of section (Pl. 2).

ments of the Cretaceous section (Pl. 1). The thickness of this member ranges from 15 to 95 feet, but it is nevertheless readily recognized.

The Aspen shale owes its distinctiveness to the fact that it is composed of fine-grained, light-colored, siliceous material, probably water-laid volcanic ash (Pl. 4, fig. 2). In addition, it contains great numbers of teleost fish scales. A typical outcrop appears as a silvery gray ribbon midway between the Dakota (?) and Frontier sandstones where they outcrop along the south flank of the Uinta Mountains (Pl. 4, fig. 1).

Weathering of this stratum produces silver-gray, splintery, sharp-edged, fissile fragments strewn in great profusion over the underlying strata. Thus, where glacial or other gravels tend to mask the underlying rocks, the approximate position of the base of the Cretaceous may often be inferred from the presence of these fragments.

Outcrops of the Aspen shale on the north side of the Uinta Mountains have been described in many reports, probably most thoroughly by Veatch (1907, p. 64–65). There, also, it typically consists of dark-gray and black shales containing numerous fish scales. Although 2200 feet of this siliceous shale is reported in southwestern Wyoming as compared with a maximum of 97 feet in the Uinta Basin, correlation seems certain. Again, it seems there must have been a barrier in the vicinity of the Uinta Mountains during deposition of the Aspen shale similar to the one that probably existed during deposition of the Dakota (?) and Bear River formations.

In northwestern Colorado, Gale (1910, p. 61) and Sears (1924, p. 286) report a gray siliceous shale with abundant fish scales in the same position above the Dakota (?) sandstone as the Aspen occupies elsewhere. They consider this bed equivalent to the Mowry shale of the Bighorn Mountain region. The Mowry and Aspen shales are considered correlatives by Bartram (1937, p. 912) as well as by many others.

The Aspen shale is not present in Castle Valley or the adjoining Book Cliffs area of Colorado. The writer searched diligently for these siliceous fish scale-bearing beds in the vicinity of Farnham dome but found none. The logical inference is that the southern limit of the Aspen is not far south of the Uinta Mountains.

Overlying the Aspen shale and underlying the Frontier sandstone is the middle shale member, a thin clay shale with variable thickness and composition. In the western part of the Uinta Basin, along the Duchesne River, it is a gray clay shale 42–108 feet thick. It becomes thicker and darker toward the east and at the Colorado border is 98 feet thick and carbonaceous.

This shale may be equivalent to the dark fissile shale underlying the Ferron sandstone in Castle Valley described by Spieker (1931, p. 19). Sears (1924, p. 286) reported a dark platy shale in similar stratigraphic position in northwestern Colorado. Tentative correlation between this shale and the carbonaceous facies of the middle shale member of the Mancos shale, as shown in the eastern end of the Uinta Basin, seems reasonable.

Immediately above the middle shale member is a sequence of sandstone beds—the Frontier sandstone member. They are 756 feet thick along Red Creek, in the western part of the Uinta Basin, and thin steadily eastward until at the Colorado

FIGURE 1. FRONTIER SANDSTONE HOGBACK NEAR VERNAL
This hogback is nearly continuous for 50 miles.

FIGURE 2. ASPEN SHALE
Its outcrop appears as a silver-gray ribbon. It contains fish scales and is highly siliceous. Outcrop is near Jensen.

FRONTIER SANDSTONE AND ASPEN SHALE IN THE VERNAL-JENSEN AREA

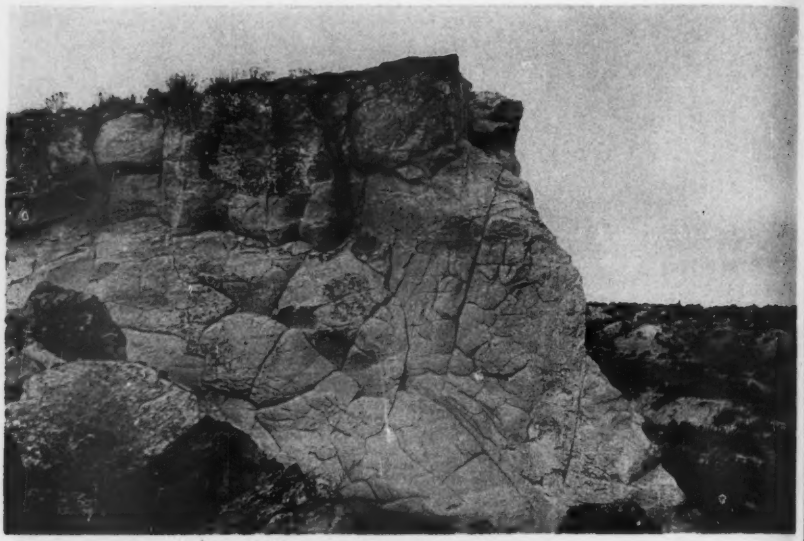


FIGURE 1. ASPHALT RIDGE SANDSTONE
Near south end of Asphalt Ridge. It consists of colorless, well-sorted quartz grains.

FIGURE 2. FRONTIER SANDSTONE NEAR VERNAL
It is composed of colorless, well-sorted quartz grains. Weathering into pillow-shaped forms is common.

ASPHALT RIDGE AND FRONTIER SANDSTONES IN VERNAL VICINITY

border they are an interbedded succession of very thin sandstones and sandy shale 163 feet thick. Because of their lithologic character, fauna, and stratigraphic position, they are correlated with the Frontier sandstone of southwestern Wyoming.

Wherever exposed, this sandstone member of the Mancos shale forms a prominent hogback (Pl. 4, fig. 1). The uppermost part is often very white and makes an easily recognized outcrop. (Pl. 5, fig. 2.) Coal beds of commercial value occur in the Frontier at Deep Creek and Ashley Valley.

The following section shows the character and thickness of this sandstone in the west end of the Uinta Basin:

Section (1) of the Frontier sandstone at Red Creek

Mancos shale, upper shale member, drab clay shale.

Frontier sandstone	Feet
Sandstone, very white, massive	15
Sandstone, brown, hard	10
Shale, sandy, drab (Tongue of Mancos shale)	30
Sandstone, white, prominent ledge maker	17
Shale, sandy, drab (Tongue of Mancos shale)	10
Coal, well decomposed	2
Shale, underclay	16
Sandstone, brown, thin-bedded	8
Shale, gray, calcareous (Tongue of Mancos shale)	17
Sandstone, thin-bedded, ledge maker	2
Sandstone, brown, white, massive	82
Shale, carbonaceous, some thin coal seams (Probably a tongue of Mancos shale)	56
Sandstone, white, brown, massive	127
Shale, gray, sandy	18
Sandstone, white, brown, massive, coarse, arkosic	95
Shale, gray, sandy	26
Sandstone, buff, massive, cliff forming	47
Sandstone, light brown, soft, massive, has several oyster beds	50
Shale, sandy, mostly covered	116
Sandstone, thin-bedded, oyster bed	12
Total	756

Mancos shale, middle shale member, dark gray.

In this section, tongues of Mancos shale, although thin, are evident (Pl. 1). Moreover, it is also apparent that the upper part of the Frontier sandstone at Red Creek is equivalent stratigraphically to the lower part of the upper shale member of the Mancos farther east along Duchesne River and Rock Creek. Still farther east at Vernal, the Frontier has thinned to 266 feet, of which only half is sandstone, the remainder being sandy shale and coal. Near the Colorado border these sandstone strata become little more than a gray sandy shale 163 feet thick, throughout which beds of hard thin-bedded calcareous brown sandstone less than 2 feet thick are scattered. Nevertheless, this sandy layer is harder than the surrounding shales and forms a rounded hogback from which no hard layers protrude (Pl. 10, fig. 1).

The lower boundary of the Frontier sandstone is marked by a thin sandstone bed or by the bottom of a zone of sandy shale alternating with thin beds of sandstone. Plate 1 indicates extensive continuity of this lower boundary, but the lower bed may not be the same everywhere. Exposures are somewhat intermittent.

The upper boundary shows a definite intertonguing of the main body of the Mancos shale with the top beds of the sandstone. This intertonguing causes the

contact between the top of the Frontier sandstone and the overlying main body of Mancos shale to descend about 300 feet between Red Creek and the Duchesne River (Pl. 1).

The thinning of the Frontier sandstone is due to two factors: intertonguing and change of facies. The former is most important in the Tabby Mountain vicinity but farther east is of little consequence. Since the shore line of the Upper Cretaceous sea described by Spieker and Reeside (1926, p. 429–438) had a general north-south direction, a natural gradual change of facies is to be expected toward the east. This can be seen by following any of the upper beds of the Frontier sandstone eastward for they progressively become finer-grained, more thinly bedded, more calcareous and argillaceous, and finally disappear.

The following fossils were collected by the writer from the Frontier sandstone at Red Creek and were identified by Dr. John B. Reeside, Jr., of the U. S. Geological Survey:

C. cf. *C. perangulata* Whiteaves
C. cf. *C. undifera* Meek
Brachydontes multilingerus Meek
Panope cf. *P. simulatrix* Whiteaves
Inoceramus aff. *I. labiatus* Schlotheim
Ostrea soleniscus Meek
Corbula cf. *C. subtrigonalis* Meek and Hayden

The first of these forms is of Greenhorn age, though the remainder are longer-ranging species. Reeside (personal communication) reports that Lupton collected fossils of Carlile age from the lower third of the Mancos formation at Duchesne River, which would be in the same stratigraphic zone (Pl. 1).

The Frontier formation of southwestern Wyoming, as described by Veatch (1907, p. 65–69), is considered middle and upper Benton in age. *Inoceramus labiatus* Schlotheim, *Ostrea soleniscus*, and numerous undescribed species of *Corbula* were collected from it by T. W. Stanton. A direct correlation of at least a portion of this formation with the Frontier sandstone of the Uinta Basin area seems assured.

The Ferron sandstone of the Book Cliffs and Wasatch Plateau region contains Carlile fossils, according to Clark (1928, p. 13–14). A tentative correlation therefore can thus be made between the upper part of the Frontier sandstone of the Duchesne River area and the Ferron sandstone.

The fossils collected in the Frontier at Red Creek are mainly a brackish-water fauna. The presence of coal beds in the section is further evidence that the sandstones were deposited on a coastal plain where occasional vegetation flourished. A marine fauna in the top of the section shows that the sea came in during the later part of the depositional history.

Oscillations of the shore line occurred at several times. Regressions of the sea produced the eastward-thinning tongues of sandstone, and transgressions formed the westward-projecting tongues of Mancos shale (Pl. 1).

The upper shale member, or main body of the Mancos shale, overlying and intertonguing with the Frontier sandstone, is lithologically a simple unit. It consists of marine gray to drab clay shale and appears to be homogeneous throughout the Uinta Basin. This appearance may be deceptive, for a heavy mantle of clay soil

covers it everywhere. The best exposures in the Uinta Basin are below Jensen where the Green River has removed part of the soil mantle. This section is as follows:

Section (10) of Mancos upper shale member at the Green River

Asphalt Ridge sandstone, buff, white grains of quartz.

Upper shale member of the Mancos shale	Feet
Shale, drab, gray, yellow, sandy, clay with many beds of brown, hard, fine-grained, calcareous sandstone, none thicker than 7 feet	747
Shale, yellowish, gray clay, slightly sandy	2360
Shale, drab, clay	730
Shale, bluish gray, clay	785
Shale, dark blue, clay	450
Total	5072

Frontier sandstone member, white, brown, massive.

The hard, brown, thin sandstone beds in the upper part of the above sequence are probably the feather edges of larger sandstone bodies that thin out eastward into the main body of the Mancos shale (Pl. 1).

AGE AND CORRELATION: The Mancos shale in the eastern part of the Uinta Basin was mapped by Gale (1910, p. 61), who traced it there from its known localities in Colorado. Where it is described by Erdmann (1934, p. 29–31) in the Book Cliffs area of the Utah-Colorado border, it is equivalent in age to the Benton and Niobrara of the Colorado group and to beds up to and including the type Lewis of the Montana group. The Mancos in the vicinity of Vernal is equivalent to the Benton and Niobrara plus an unknown portion of the Montana group. An exact correlation cannot be made since the fossils collected by Tolmachoff from the overlying Mesaverde beds may only be referred to as post-Niobraran in age, according to Reeside (personal communication). In the vicinity of Tabby Mountain, however, the beds mapped and described as Mancos are below the top of the Niobrara faunal zone and therefore represent only the middle Mancos of the Wasatch Plateau-Book Cliffs region described by Spieker (1937, p. 16–21) (Pl. 3).

This upper shale member of the Mancos in the Vernal area is probably correlative with some part of the Hilliard shale, described by Veatch (1907, p. 69–72) in southwestern Wyoming on the north side of the Uinta Mountains. Both occupy the same general stratigraphic and faunal zones, and their lithologic similarity is striking.

MESAVERDE GROUP

General statement.—For mapping purposes, any sequence of sandstone beds conformably overlying the typical drab Mancos shale has been called the Mesaverde Group Undivided. Actually, the beds mapped differ considerably in age (Pls. 1, 2). Nevertheless, these strata, no matter what age, may generally be divided into two parts: a lower marine succession of sandstones, and an upper of brackish-water sandstones, sandy shales, carbonaceous shales, and coal.

These Mesaverde strata are exposed in two distinct areas in the Uinta Basin: the Tabby Mountain area described by Lupton (1912, p. 579–594) in 1912, and the Vernal region by Gale (1910, p. 63–71) in 1910. For greater facility and simplification in description these areas will be discussed separately.

Tabby Mountain area.—The strata most typically representative of the Mesaverde

group as it occurs in the western part of the Uinta Basin are exposed along Red Creek and the Duchesne River on the west and east sides of Tabby Mountain, respectively.

GENERAL FEATURES: The rocks classed as Mesaverde in this area outcrop intermittently from Currant Creek to a few miles east of the Duchesne River (Pl. 2). On the basis of their most complete exposure at Red Creek they may be divided into two parts: a lower marine series represented by 869 feet of massive sandstone, and an upper fresh- and brackish-water series 2132 feet thick composed of sandstones, sandy shales, and coal.

The upper boundary of the Mesaverde group, throughout the area, is unconformable with a succession of conglomerates, continental sandstones, and shales for which the name Currant Creek formation is proposed. The lower boundary is the main body of the Mancos shale, the contact being represented by transition and intertonguing.

LITHOLOGIC CHARACTER AND THICKNESS: The best exposure and the maximum thickness of the Mesaverde group of strata are along Red Creek, (Pl. 6), as shown in the following section:

Section (1) of the Mesaverde Group at Red Creek

Currant Creek formation, conglomerate, pebbles of gray quartzite and black chert underlaid by "Transition beds".

Unconformity, exact contact undeterminable.

Mesaverde group	Feet
Sandstone, gray, buff, soft; shale; coal	69
Sandstone, erratic bedding, conglomeratic	32
Sandstone, gray, buff; shale, sandy, partly covered, contains one large, many small coal beds	452
Sandstone, gray, soft, some thin hard brown beds; shale, gray, sandy	144
Sandstone, calcareous, Fossil Zone #5	3
Sandstone, white, gray, cross-bedded, sugary	121
Covered, probably soft sandstone and shale	31
Sandstone, soft, thin-bedded, coal seam	91
Sandstone, gray, soft, massive, sand concretions	236
Sandstone, fine grain, thin-bedded, gray	18
Sandstone, gray, cross-bedded	92
Covered, gully and spring	41
Sandstone, gray, thin-bedded, ledge maker	65
Sandstone, brown, yellow, ripple marks	82
Sandstone, gray, soft, interbedded with shale	61
Sandstone, exceedingly white, fine-grained	4
Sandstone, gray, buff, massive, sugary	113
Covered, probably sandy shale	142
Sandstone, white, massive, very prominent	91
Sandstone, buff, cross-bedded	53
Shale, gray, brown, sandy	54
Sandstone, gray, massive, prominent	48
Shale, gray, brown, sandy	12
Sandstone, gray, soft; shale, gray, sandy	61
Sandstone, brown, hard, ledge forming	9
Shale, gray, calcareous	7
Limestone, sandy, Limestone Fossil Zone	2
Shale, dark gray, sandy	44
Limestone, coquina, Limestone Fossil Zone	2
Sandstone, white, distinctive	37
Covered, probably shale	24
Sandstone, hard, calcareous, Fossil Zone #3	48
Shale, black, carbonaceous	2
Sandstone, gray, sugary, Fossil Zone #2	88

Looking west from Tabby Mountain. Note unconformities at Uinta (?)-Currant Creek and Currant Creek-Mesaverde contacts. Note also thickening down dip of strata in Currant Creek formation.

CRETACEOUS SECTION ALONG RED CREEK

Note overlap of Wasatch formation by Green River shale.

LOOKING EAST ALONG THE RIM ROCK HOGBACK

Sandstone, hard, prominent ledge maker	84
Sandstone, white, gray, cross-bedded	149
Sandstone, brown, cross-bedded	144
Sandstone, gray, fine-grained, very soft	139
Sandstone, gray, massive, Fossil Zone #1	106
Total	3001

Mancos shale, upper shale member, drab shale.

The fossils found in this section at the various Fossil zones show that the strata below the Limestone zones are dominantly marine, whereas those above are of brackish- and fresh-water deposition.

Along the Duchesne River, the Mesaverde group is again well exposed, but only the basal portion of the above section is present (Pl. 1). Farther east, on Farm Creek, a somewhat greater thickness is present because the position of the unconformity in the section fluctuates, but still farther east, at Rock Creek, both the Currant Creek and Mesaverde are completely covered by the Uinta (?) formation (Pl. 2).

STRATIGRAPHIC RELATIONS: It is apparent from Plate 1 that the shore line of the sea in which the Mesaverde strata were deposited was roughly north-south. Thus a natural change of facies going seaward toward the east accounts for a considerable part of the loss of sandstone in the section between Red Creek and Duchesne River. Eastward wedging out of the lower sandstone beds may account for part of this loss. Continuous exposures of any of the beds of the Mesaverde group are lacking, hence the nature of the exact stratigraphic change is obscure.

The Mesaverde group is overlain unconformably by a prominent succession of conglomerates, sandstones, and variegated shales of continental origin for which the name Currant Creek is proposed. The unconformity is of considerable magnitude and represents a period of uplift and erosion before deposition of the Currant Creek formation. In spite of the relief along the unconformity (Pl. 1), there is little discordance in dip above and below the contact. The exact stratigraphic position of the unconformity is indefinite because of the transitional nature of the section at the contact. This condition is shown along the Duchesne River in the following section:

Detailed section (2) of the Mesaverde group upper boundary at Duchesne River

Currant Creek formation, definitely identified; conglomerate, large quartzite and black chert pebbles, a prominent ledge former 100 feet thick.

Transitional strata	Feet
1. Sandstone, white, massive, ledge former, coarse; contains black chert pebbles at top in many small lenses	145
2. Sandstone, reddish brown, coarse-grained, some limonitic material	13
3. Sandstone, cross-bedded, coarse-grained; black chert pebbles in small lenses	10
Unconformity (?)	
4. Sandstone, white, alternating with red shale	83

Mesaverde group, definitely identified; sandstone, alternately hard and soft, well sorted, typically marine, 69 feet thick.

In this section, the writer places the unconformity between members No. 3 and 4 because of the black chert pebbles in No. 3. No chert pebbles were observed any-

where in the 3001 feet of the Mesaverde group at Red Creek except in the transition zone at the base of the Currant Creek formation. The logical inference is that these pebbles mark the unconformity. Such a transitional sequence must have resulted from considerable reworking of the strata of the Mesaverde group over the erosion surface upon which the Currant Creek formation was ultimately deposited.

AGE AND CORRELATION: The following fossils were collected by the writer from the Mesaverde section at several localities and were identified by Dr. John B. Reeside, Jr., of the U. S. Geological Survey.

Fossils from the Mesaverde group at Red Creek

Fossil Zone No. 1	
Tellina? sp.	Marine species
Mactra? sp.	Marine species
Fossil Zone No. 2	
Cardium aff. *C. curtum* Meek and Hayden.	Marine species
Fossil Zone No. 3	
Ostrea soleniscus Meek	Brackish-water species
Brachydontes multilinigerus Meek	Brackish-water species
Cardium cf. *C. pauperculum* Meek	Marine species
Tellina? sp.	Marine species
Limestone Fossil Zones	
Ostrea soleniscus Meek	Brackish-water species
Brachydontes multilinigerus Meek	Brackish-water species
Corbula cf. *C. subtrigonalis* Meek	Brackish-water species
Corbula cf. *C. undifera* Meek	Brackish-water species
Panope simulatrix Whiteaves	Brackish-water species
Melania cf. *M. insculpta* Meek	Brackish-water species
Anomia sp., simple form	Marine species
Lucina? sp.	Marine species
Cardium pauperculum Meek	Marine species
Fusus? sp.	Marine species
Pseudocolumna? sp.	Fresh-water species
Goniobasis? sp.	Fresh-water species
Planorbis sp. small form	Fresh-water species
Fossil Zone No. 5	
Ostrea soleniscus Meek	Brackish-water species
Anomia sp. simple form	Brackish-water species

At Duchesne River and Farm Creek the following fossils were collected from the same general zone as the Fossil Zone No. 2 along Red Creek:

Idonearca ("*Cucullaea*") sp. undescribed	Marine species
Ostrea sp.	Marine species
Anomia sp. simple form	Marine species
Astarte? sp.	Marine species
Cardium pauperculum Meek	Marine species
Tellina? sp.	Marine species
Legumen? sp.	Marine species
Exogyra suborbiculata Lamarck, of Stanton	Marine species
Gyrodes aff. *G. conradi* Meek	Marine species

This group of fossils, according to Reeside (personal communication), contains no forms younger than the Niobraran faunal zone. The lower 2304 feet of the Mesaverde group of the Tabby Mountain area lie below the top of that zone. The age of the upper 697 feet is as yet unknown. Lupton's collections from the upper

part of the Mancos shale in the vicinity of Duchesne River are of Niobraran age, according to Reeside. It seems certain, therefore, that the greater part of the Mesaverde group in this area is Niobraran and belongs to the Colorado division of the Upper Cretaceous (Pl. 1).

The Niobraran Mesaverde sandstones of the Tabby Mountain area are contemporaneous with the middle shale member of the Mancos in the Wasatch Plateau-Book Cliffs region, which shale, according to Spieker (1931, p. 16), contains the same faunal zone. The change in lithologic facies is illustrated in Plate 3.

According to Veatch (1907, p. 69–72), the lower half of the Hilliard shale on the north side of the Uinta Mountains is mainly of Niobrara age and thus by faunal correlation is roughly equivalent to both the Mesaverde sandstones and a portion of the Mancos shale of the Tabby Mountain area (Pl. 3).

Fossils collected by Tolmachoff (1942, p. 41–60) from Asphalt Ridge in the Vernal region show that the sandstones of the Mesaverde group there are above the Niobrara faunal zone. It follows, then, that the sequence of beds designated Mesaverde in the Tabby Mountain area is equivalent stratigraphically and faunally to the main body of the Mancos shale in the vicinity of Vernal (Pl. 1).

CONDITIONS OF ORIGIN: The sandstone of the Mesaverde group below the limestone fossil zones was deposited on the beach and near the shore of the sea in which, farther east, the muds and clays now forming the main body of the Mancos shale in the Vernal region were laid down. A gradual change of facies is noted from west to east. Probably there were fluctuations of the shore line as intertonguing seems to be present in the lower part of the sandstone section. This relationship would no doubt be more obvious if the great overlap of the Tertiary rocks between Duchesne River and Asphalt Ridge were not present.

With the retreat of the sea, following a brief period of limestone deposition, sediments overlying the limestone fossil zones were deposited in a near-shore brackish-water environment that occasionally was flooded by fresh water from near-by streams. Thus both brackish- and fresh-water faunas are present. Near the end of this depositional period, extensive swamps were present, and the coal beds of the upper part of the section were formed.

Vernal region.—Strata typical of the Mesaverde group of the eastern part of the Uinta Basin outcrop in this general area, particularly to the east of Vernal where they may be studied most advantageously.

GENERAL FEATURES: The Mesaverde strata are present in this area from Asphalt Ridge to the Colorado border (Pl. 2). The basal marine beds form a prominent hogback called the "Rim Rock" that extends continuously from the Green River to the Rangely anticline. Throughout this area, the Mesaverde is overlain unconformably by Tertiary formations. This unconformity, which undulates to some extent, rises stratigraphically from west to east about 3000 feet between Asphalt Ridge and the Colorado border (Pl. 1).

In Asphalt Ridge, the lithologic characteristics of the Mesaverde sandstones are somewhat masked because the rock is frequently saturated with bituminous material. This saturation is definitely related to the unconformity between the Tertiary and Upper Cretaceous as it occurs in the beds both above and below the contact. More-

over, the unconformity fluctuates stratigraphically in the Mesaverde section, and nearly always the beds most closely associated with it have the greatest bituminous content. Spieker (1930, p. 77–98) described these saturated sandstones.

LITHOLOGIC CHARACTER: The Mesaverde group of this area consists of two distinctive parts. The lower contains typically marine littoral sandstones, and the upper consists of brackish-water sandstones interbedded with sandy red, pink, and reddish-purple shales and thin coal seams. Gale (1910, p. 1–265) mapped these beds in 1907 as lower and upper Mesaverde formation. The present work shows that the basal marine sandstones may be divided into two easily recognized members which are tongues of sandstone separated by shale. Each member thins toward the east and eventually disappears into the main body of the Mancos shale.

The upper, or brackish-water, portion is not continuously enough exposed to allow lithologic division. At the Rangely anticline, however, it contains several prominent beds which might serve as marker horizons.

For convenience in description, the Mesaverde is subdivided into the following, which have been given those geographical names best describing their occurrence:

Williams Fork formation (includes more than type Williams Fork)
Rim Rock sandstone
Asphalt Ridge sandstone

The subdivisions are discussed in order of deposition.

Asphalt Ridge sandstone.—This lowermost formation of the Mesaverde group may be studied to best advantage along Asphalt Ridge, near Vernal.

GENERAL FEATURES: The name Asphalt Ridge is here proposed for the lower of the two basal marine sandstones of the Mesaverde group in the Vernal region. This member is exposed most typically in Asphalt Ridge. It is usually a single, massive, yellow to white soft sandstone bed, about 100 feet thick, bounded on top by a thin tongue of Mancos shale and on the bottom by the main body of the Mancos shale.

The stratum is unique in that nowhere in Asphalt Ridge does it contain bituminous material, whereas sandstones both above and below are saturated.

The member may be traced, with interruptions, from the north end of Asphalt Ridge about 15 miles to the Rim Rock at Green River, where it feathers out into the main body of the Mancos shale.

LITHOLOGIC CHARACTER AND THICKNESS: The following section at the north end of Asphalt Ridge shows the nature and relations of this member:

Section (7) of Asphalt Ridge sandstone at north end of Asphalt Ridge

Shale, sandy, gypsiferous, probably tongue of Mancos shale.

Asphalt Ridge sandstone	Feet
Sandstone, soft, massive, yellow, no partings, no bituminous material, weathers into smooth face	101
Sandstone, thin-bedded, considerable bituminous material, argillaceous, limonitic in places: shale, sandy, probably represents tongue of the Mancos shale.	

This is not a typical exposure for in all other places the main body of the Mancos shale immediately underlies the Asphalt Ridge sandstone. Here, however, several

sandstone beds lie below the thin-bedded argillaceous sandstone shown in the above section. They cannot be traced from this locality (Pl. 1) and probably represent local sand lenses in the top of the Mancos shale.

Farther south the Asphalt Ridge sandstone becomes almost pure white, but the other characteristics remain essentially the same (Pl. 5, fig. 1). Still farther south and east at the Green River the member is but 51 feet thick, and it disappears in the Mancos shale within a short distance along the Rim Rock.

STRATIGRAPHIC RELATIONS: The eastward thinning of the Asphalt Ridge sandstone is mostly due to a gradual lateral change of facies such as is natural in going seaward. The actual contact between the Asphalt Ridge sandstone and the underlying Mancos shale ranges through a considerable vertical distance because of the transitional nature of the zone separating the two.

The upper boundary of the Asphalt Ridge sandstone is more or less transitional also, for it involves only the passage from marine sandy shale to soft massive sandstone. In some places the transition is so gradual that it is impossible to draw a sharp line between the Asphalt Ridge and the overlying Rim Rock sandstone.

AGE AND CORRELATION: Tolmachoff (1942, p. 41–60) collected and identified the following fossils from the Asphalt Ridge sandstone at the north end of Asphalt Ridge:

Enchodus sp.
Apateodus sp.
Scapanorhynchus sp.
Placenticeras meeki J. Boehm
Turritella sp.
Gyrodes sp.
Margarita sp.
Fissurella sp.
Leptosolen conradi Meek
Anatina lineata Stanton
Cymella bella Conrad
Lucina sp.
Cardium kayi sp. nov.
Tellina sp.
Cymbopora sp.
Yoldia evansi Meek and Hayden
Crenella burkei sp. nov.

This assemblage of fossils can be considered only as post-Niobraran according to Reeside (personal communication).

CONDITIONS OF ORIGIN: The Asphalt Ridge sandstone was deposited near the shore of the same sea in which the muds and clays of the Mancos shale were laid down. A slight oscillation of sea level changed the positions of the various environments of deposition long enough for a tongue of sand to be deposited in the zone of mud and fine sand deposition. This change of position was no doubt accomplished slowly, allowing deposition of the transitional zones above and below the sandstone tongue.

Rim Rock sandstone.—This prominent member of the Mesaverde group is present as a mappable lithologic unit from Vernal eastward to the Colorado border. It may be studied advantageously anywhere along its outcrop.

GENERAL FEATURES: The upper member of the marine basal portion of the Mesaverde group is the most distinctive sandstone sequence in that whole group of rocks. It outcrops continuously from Asphalt Ridge to Rio Blanco County, Colorado, as a hogback ridge called the "Rim Rock". For this reason the writer proposes to name this prominent member the Rim Rock sandstone (Pl. 7).

The lower boundary of the Rim Rock sandstone in the western part of the area

is the thin tongue of Mancos shale separating it from the underlying Asphalt Ridge sandstone. Farther east, this thin shale passes into the main body of the Mancos where the Asphalt Ridge disappears (Pl. 1). The upper boundary, at the base of the Williams Fork formation (includes more than type Williams Fork), is usually marked by a green or gray shale overlain by brackish-water, brown, thin-bedded sandstones interbedded with pink, purple, white, and black shale and thin coal beds.

Throughout a portion of its outcrop in the Asphalt Ridge vicinity the Rim Rock sandstone is impregnated with varying amounts of bituminous material. At the north end of Asphalt Ridge it is of commercial value.

LITHOLOGIC CHARACTER AND THICKNESS: The Rim Rock sandstone is essentially a simple lithologic unit, particularly at its type locality in the Rim Rock hogback. There, it is a gray, medium- to fine-grained sandstone, cross-bedded in some places, thinly bedded in others, and containing a few beds of shale that are tongues from the main body of the Mancos shale. Varying amounts of black chert grains impart a speckled appearance to the rock. The following section shows the character and thickness of the Rim Rock sandstone at its type locality:

Section (11) of the Rim Rock sandstone at the Rim Rock, 9 miles east of the Green River

Williams Fork formation (includes more than type Williams Fork); buff and brown sandstones interbedded with green, gray, pink, and purple shales. Shale, gray, sandy.

Rim Rock sandstone	Feet
Sandstone, white, soft, massive	48
Sandstone, greenish yellow, distinctive	2
Shale, gray and brown, sandy, probably tongue of Mancos shale	33
Sandstone, fine-grained, gray, speckled	127
Sandstone, thin-bedded, speckled	36
Shale, light gray, sandy	4
Sandstone, cross-bedded, gray	63
Sandstone, brown, some shale	12
Shale, gray, sandy, tongue of Mancos shale	36
Sandstone, fine-grained, speckled	41
Sandstone, buff, soft, massive	14
Total	416

Mancos shale, main body, sandy in upper part.

Like the Asphalt Ridge, the Rim Rock sandstone thins eastward, taking on more and more shale, until over half the sequence is argillaceous, as shown in the following section at the Rangely anticline (Pl. 1):

Section of the Rim Rock sandstone at Rangely Anticline

Williams Fork formation (includes more than type Williams Fork); gray, brown, pink, purple, and black shales with prominent brown and white massive sandstones at frequent intervals.

Shale, dark, sandy, with thin beds of brown sandstone.

Rim Rock sandstone	Feet
Sandstone, gray, soft, sugary, thin-bedded	87
Shale, drab, gray, sandy, with numerous layers of buff, hard sandstone, none thicker than 5 feet; probably tongue of Mancos shale	322
Sandstone, white, massive, forms the rim rock of the Rangely anticline	62
Shale, drab, sandy, tongue of Mancos shale	46
Sandstone, drab, thin-bedded, argillaceous	8
Shale, drab, tongue of the Mancos shale	23
Sandstone, buff, thin-bedded, concretions	32
Total	580

Mancos shale, main body, drab shale.

FIGURE 1. CONGLOMERATE BED IN CURRANT CREEK FORMATION AT THE DUCHESNE RIVER
It is composed of pebbles of black chert and gray quartzite.

FIGURE 2. RIM ROCK SANDSTONE AT RANGELY ANTICLINE
Lower stratum tongues out a short distance to the east.

CURRANT CREEK CONGLOMERATE ALONG DUCHESNE RIVER AND RIM ROCK SANDSTONE AT RANGELY ANTICLINE

Looking east to Farm Creek peak. All formations dip south.

UNCONFORMITIES ALONG THE DUCHESNE RIVER

This section shows that the Rim Rock sandstone is split by three shale beds, the lower and middle being continuous with the main body of the Mancos shale a few miles to the east.

The upper and much thicker shale bed could not be traced due to insufficient exposures. The portion of the Rim Rock sandstone forming the rim of Rangely anticline is conspicuous by its light-gray color and the fact that it weathers in characteristic rounded forms at its top (Pl. 8, fig. 1).

One of the more interesting exposures of the Rim Rock sandstone is in the north end of Asphalt Ridge where its basal portion forms the upper "Asphalt bed" in local nomenclature. Here, the basal conglomerate of the Tertiary Duchesne River formation unconformably overlies the Rim Rock sandstone as shown in the following section:

Section (7) of the Rim Rock sandstone at the north end of Asphalt Ridge

Duchesne River formation, reddish-orange shales, buff sandstones and conglomerate.
Basal conglomerate.

Unconformity

Rim Rock sandstone	Feet
Sandstone, fine-grained, saturated with bituminous material, upper "Asphalt bed"	28
Covered, probably soft sandstone	27
Sandstone, thin-bedded, partly saturated with bituminous material	6
Sandstone, coarse, erratically bedded, highly saturated with bituminous material	51
Total	112

Shale, sandy, lime concretions, gypsum, tongue of Mancos shale.

The unconformity in this section is of considerable magnitude, for part of the Rim Rock sandstone, all the Williams Fork formation, the Green River formation, and the Uinta formation are missing.

STRATIGRAPHIC RELATIONS: The intertonguing of the Rim Rock sandstone and the Mancos shale is shown on Plate 1. This relationship causes the sandstone-shale contact to migrate upward, proceeding eastward. The same relation has been noted elsewhere between the Mesaverde sandstones and the Mancos shale, particularly in the Book Cliffs area of Utah and Colorado where the upward migration of the sandstone-shale contact is measured in thousands of feet. The stratigraphic rise in contact between the Mancos shale and the Rim Rock sandstone is much less, in the Uinta Basin, probably not more than 100 feet.

The top of this member is marked by a sharp change in lithologic character. The overlying varicolored shales and buff sandstones of the Williams Fork strata make a sharp boundary with the gray marine Rim Rock sandstone.

AGE AND CORRELATION: No fossils were collected from the Rim Rock sandstone. The fossils in the Asphalt Ridge sandstone described by Tolmachoff (1942, p. 41–60) belong to a post-Niobrara faunal zone. Inasmuch as these two members are very close stratigraphically, the Rim Rock may also be considered post-Niobrara.

The Rim Rock sandstone forms the rim rock of the Rangely anticline and makes a prominent escarpment throughout the Raven Park area. Spieker (personal communication) and Reeside traced the strata of the Book Cliffs region, described by

Fisher (1936, p. 15–16) and Erdmann (1934, p. 34–36), into the Raven Park area and found that the rim rock there is about 1500 feet lower stratigraphically than the base of the Sego sandstone of Erdmann's section. This places the Rim Rock stratigraphically at about the position of the middle of the Star Point sandstone of the Book Cliffs.

According to Hancock (1925, p. 15–18) and Sears (1924, p. 288–290), a sandstone similar in lithologic character and stratigraphic position to the Rim Rock sandstone forms the rim of the Axial Basin. Hancock considered this bed the base of his Iles formation, which constitutes the lower marine portion of the Mesaverde group. A tentative correlation seems possible, therefore, between the basal part of the Iles formation and the Rim Rock sandstone (Pl. 3).

North of the Uinta Mountains, at the base of the Adaville formation which, according to Reeside (personal communication), occupies the same general faunal zone as the Mesaverde group in central Utah, there is a 200-foot white prominent sandstone locally called the Lazeart (Veatch, 1907, p. 72–73). The writer has noted several times the great similarity between this bed and the Rim Rock sandstone and suggests that the two may be correlatives.

CONDITIONS OF ORIGIN: The Rim Rock sandstone was deposited on the beach or near the shore of the Upper Cretaceous sea. The expanse and general continuity of this sandstone, as well as cross-bedding, indicates that strong currents were active during its deposition, for they spread rather widely the sand of which it is composed.

The angularity and general coarseness of the sand grains, together with many chert and occasional feldspar fragments, particularly in the Asphalt Ridge vicinity, indicate that the source of this material was not very far distant.

During deposition, continued fluctuation of the strand line gave rise to intertonguing of sandstone bodies with the tongues from the main body of Mancos shale. Withdrawal of the sea is recorded by tongues of sandstone which represent an eastward migration of the near-shore conditions. Transgression of the sea caused the zone of clay and mud deposition to move westward, forming the tongues of shale.

Williams Fork formation.—This rather drab member of the Mesaverde group is as a rule poorly exposed, particularly when it dips steeply; detailed study of it is then difficult.

GENERAL FEATURES: Conformably overlying the Rim Rock sandstone is a heterogeneous sequence of strata that Gale (1910, p. 63–71) in 1907 mapped as upper Mesaverde formation. Since lithologically and stratigraphically they are approximately the equivalent of the Williams Fork formation of Moffat County, Colorado, described by Hancock (1925, p. 18–20), the author has chosen to designate them as Williams Fork. It should be noted, however, that these strata so designated include, both at their top and bottom, more than the type Williams Fork formation. Equivalents of the Iles formation at the bottom and the Lewis shale at the top are no doubt present in the Williams Fork formation of this paper (Pl. 3).

The Williams Fork formation is almost continuously exposed along the Rim Rock hogback, where it is overlain unconformably by Tertiary formations. Since it has more shale and softer sandstones than the underlying Rim Rock beds, its position is usually marked by a depression paralleling the Rim Rock hogback (Pl. 7).

LITHOLOGIC CHARACTER AND THICKNESS: A typical exposure of the Williams Fork beds exhibits gray, brown, black, pink, and reddish-purple sandy shales interbedded with numerous prominent strata of buff, brown, and white sandstone. Some coal and carbonaceous material are always present. The pink and purplish colors in the shales are due to the heat from burning of adjacent coal beds. Lithologic heterogeneity is the most pronounced feature of these strata. In general, the sandstones are much less conspicuous than those of the Rim Rock sandstone. Conversely, the shales are more prominent, being more highly colored, no doubt due to their predominant fresh- and brackish-water origin.

By far the most complete exposure of these Williams Fork strata is on the west flank of the Rangely anticline, partly in Utah, but mainly in Colorado. The following section shows the important features of this exposure:

Section of the Williams Fork formation at the Rangely Anticline

Wasatch formation, whitish-gray, pinkish, green shales interbedded with a few thin-bedded brown sandstones. Shale, gray, containing fresh-water gastropods.

Unconformity, no discordance in dips, exact position cannot be determined.

Williams Fork formation	Feet
Shale, gray, light brown, sandy, interbedded with massive buff, brown sandstone in occasional layers	1270
Sandstone, buff, massive, coarse, ledge maker	46
Sandstone, soft, thin-bedded, some interbedded shale	358
Sandstone, light buff, forms ledge	21
Shale, light gray, sandy	20
Sandstone, buff, massive, ledge maker	11
Shale, gray, light brown	90
Sandstone, brown, thin-bedded	5
Shale, light gray, sandy	16
Shale, dark brown, sandy	14
Sandstone, buff, massive	15
Shale, carbonaceous	15
Coal, not well exposed	2
Shale, light gray, sandy	10
Sandstone, massive, buff, ledge maker	16
Shale, reddish purple	8
Sandstone, brown, hard	30
Sandstone, gray, thin-bedded	13
Shale, carbonaceous, may be coal	6
Sandstone, alternating brown and white, massive at base	141
Shale, gray	12
Sandstone, buff	5
Shale, gray and black, probably some coal	16
Sandstone, brown, massive, ledge maker	31
Shale, very sandy, some sandstone beds	33
Sandstone, brown, alternating with gray shale	61
Sandstone, massive, buff, ledge maker	16
Shale, gray, sandy	11
Sandstone, gray, white, massive, cross-bedded, very prominent, may be Trout Creek sandstone of Hancock	92
Shale, alternating with brown, hard sandstone	76
Shale, alternate gray and black, sandy	52
Sandstone, white, massive, ledge maker	31
Shale, gray, sandy	26
Sandstone, brown, white, massive	15
Shale, gray	11
Sandstone, brown, hard	5
Shale, dark brown and black	14
Total	2614

Rim Rock sandstone, gray, soft, thin-bedded at top.

This section shows that the Williams Fork consists of approximately 60 per cent shale and 40 per cent sandstone. Nearly all the shale is of continental origin, although some of it is as well bedded as the marine Mancos shale.

Some of the sandstones of the section show some resemblance to the Rim Rock sandstone. Many appear to be marine, particularly the massive, cross-bedded, prominent gray bed about 800 feet above the lowest member of the Rim Rock sandstone (Pl. 1). It has the same lithologic character and approximate stratigraphic position as the Trout Creek sandstone of the Axial Basin region, described by Hancock (1925, p. 18–20). The sandstones associated with the carbonaceous shales are of fresh- or brackish-water origin and differ considerably in appearance from the above-described marine sandstone. They are always browner and coarser and commonly contain particles of carbonaceous material.

None of the beds of the Williams Fork formation could be traced westward from the Rangely anticline. The flat dips at the west side of the structure form a depression between Raven Ridge, a cuesta of the Wasatch formation, and the Rim Rock hogback. The soil filling of this depression masks all the stratigraphy above the Rim Rock sandstone (Pl. 7). Still farther west, overlapping Wasatch and Green River formations cover more than half the strata exposed at the Rangely anticline. Much farther west, however, at the lower middle part of Asphalt Ridge, some conspicuous beds closely resemble the Trout Creek (?) sandstone of the Raven Park area (Pl. 1, Sec. 8).

STRATIGRAPHIC RELATIONS: Beginning at Asphalt Ridge and proceeding eastward, the Williams Fork formation is successively overlain unconformably by the Duchesne River, the Green River, and the Wasatch formations (Pls. 1, 2). The several regional orogenies represented by these unconformities will be discussed later.

The Williams Fork beds rest conformably on the Rim Rock sandstone. The contact represents the change from marine to brackish-water deposition.

AGE AND CORRELATION: Since Spieker's (personal communication) and Reeside's work shows a correlation between the Trout Creek sandstone and the Castlegate sandstone of the Wasatch Plateau-Book Cliffs area, the Williams Fork strata of this paper occupy generally the stratigraphic position of the Blackhawk and Price River formations of that area. Fisher (1936, p. 15–16) collected Lewis fossils from the Buck tongue of the Mancos shale and the Sego sandstone of the Price River formation, hence most of the Price River is younger than type Lewis and is Kirtland-Fruitland (Spieker and Reeside, 1925, p. 447) in age. According to Spieker, the "Lewis" shale, which immediately overlies the Williams Fork formation in the Axial Basin region, is younger than type Lewis, being approximately Fox Hills in age. It appears, therefore, that only the Blackhawk formation and the lower lower portion of the Price River formation may be correlated with the Williams Fork strata of this paper. This correlation is supported by the fact that lithologically the Williams Fork is similar to that of the Blackhawk coal-bearing formation.

The Williams Fork of this report is generally the same as the Williams Fork formation of Hancock (1925, p. 18–20), which is exposed in northwestern Colorado. The lower boundary of the Williams Fork is at the top of the Trout Creek sandstone.

This sandstone marks the base of the workable coals. A bed similar in lithologic character and in general stratigraphic position was noted at the Rangely anticline and has been described above. If this white sandstone is Trout Creek, then the base of the Williams Fork formation in the Rangely area is about 800 feet above the lowest member of the Rim Rock sandstone, or 322 feet higher than the base taken by the writer for his Williams Fork strata.

The top of the Williams Fork formation cannot be correlated with the upper boundary of the Williams Fork of this paper. Their relationship cannot be observed due to unconformable overlying Tertiary formations.

The Adaville formation, described by Veatch (1907, p. 71–76), of southwestern Wyoming occupies nearly the same stratigraphic position as the Williams Fork strata of the Uinta Basin. Veatch's fossil collections from the Adaville were identified by T. W. Stanton as upper Montana in age. Reeside (personal communication) places the Adaville in the same general faunal zone as the Blackhawk and lower Price River of the Book Cliffs region—that is, upper Pierre. Regional correlation of these coal-bearing beds seems reasonable.

CONDITIONS OF ORIGIN: The sediments forming the Williams Fork rocks were probably deposited on a broad coastal plain near sea level. Brackish-water conditions prevailed most of the time, but the sea apparently transgressed occasionally to deposit the marine sands of the section. Extensive swamps were probably present at various times as indicated by the carbonaceous shales and coal seams typical of the Williams Fork sequence.

CRETACEOUS (?) SYSTEM

CURRANT CREEK FORMATION

GENERAL FEATURES: The name Currant Creek is here proposed for the sequence of conglomerates, sandstones, and variegated shales which, in the western part of the Uinta Basin, transect the Mesaverde Niobraran beds and unconformably underlie the Eocene strata of probable Uinta age. It may be that the Currant Creek is only a facies of known and described formations, and it is hoped that further work by Dr. A. A. Baker (personal communication) will determine if this is so.

The Currant Creek formation was deposited on an erosional surface which truncates the Mesaverde group. It is exposed in various places from Currant Creek to a few miles east of Duchesne River (Pl. 2). Thick beds of conglomerate lie in the basal part, particularly along Currant and Red creeks. They become thinner and less conglomeratic toward the east, and along Duchesne River are represented by coarse sandstones containing lenses of pebbles (Pl. 8, fig. 2).

The Currant Creek beds are overlapped by probable Eocene sediments which J. L. Kay (personal communication) calls the Uinta formation, and the writer, Uinta (?). The boundary between these two formations can be seen from a distance (Pl. 6), but at closer range it is exceedingly difficult to locate exactly.

LITHOLOGIC CHARACTER AND THICKNESS: One of the best exposures of the Currant Creek is along Red Creek where the basal beds dip concordantly with the underlying Mesaverde at a steep angle (Pl. 6). The most striking feature of the formation

here, as shown in the following section, is the great thickness of conglomerate at the base:

Section (1) of the Currant Creek formation along Red Creek

Uinta (?) formation, distinctive red beds, red shales, and buff sandstones. Red shales give color to sandstone members.

Unconformity, difficult to locate closely.

Currant Creek formation	Feet
Shale, variegated, sandy, with some white thin-bedded sandstone beds	2643
Sandstone, yellow, conglomeratic	49
Shale, variegated, sandy	61
Conglomerate and sandstone, yellow, ledge maker, prominent	72
Shale, red and white	20
Conglomerate, 14-inch quartzite pebbles	131
Shale, vermillion, some white sandstone	420
Sandstone, gray, buff, conglomeratic	62
Conglomerate, interbedded with yellow coarse sandstone, conspicuous ledges	609
Covered, probably poorly cemented conglomerate	175
Conglomerate, 10-inch quartzite pebbles	10
Covered	96
Sandstone, buff, coarse	8
Conglomerate, quartzite and black chert 1- to 4-inch pebbles	194
Total	4550

Transitional Beds. (*See* Section 2.)

Unconformity, exact location not determined.

Mesaverde group strata, buff sandstones, and sandy gray shale with coal beds.

Because the Bishop conglomerate and the Duchesne River formation cap Tabby Mountain, the Currant Creek beds cannot be traced continuously east from Red Creek to the Duchesne River. Similar conglomerate, sandstone, and shale, however, at the latter locality unconformably overlie the Mesaverde group of sandstone; with no apparent discordance of dip and underlie the big "red beds" of the Uinta (?) formation. There is no doubt that they represent the same strata as the Currant Creek formation, but the conglomerates are much thinner. Some of the coarse sandstones in the Duchesne River section probably are equivalent to the conglomerates of the Red Creek section. Four miles east of the Duchesne River, this sequence contains only a few thin beds of sandstone and conglomerate near the base. East from this point the Currant Creek beds are covered by the Uinta (?) formation and apparently never outcrop again (Pls. 1, 2).

STRATIGRAPHIC RELATIONS: Although the Currant Creek formation unconformably overlies the Mesaverde group, it shows no appreciable discordance of dip at any of these contacts. There must have been irregularities on the surface upon which the basal conglomerates were deposited (Pl. 1). Apparently this surface was nearly parallel to the strata of the Mesaverde group; otherwise, discordance of dip would have resulted. The divergence of dips on the upper and lower sides of individual strata (Pl. 6) is no doubt due to mountainward thinning and represents contemporaneous deposition and orogenic movement. The transitional nature of the Currant Creek beds is definitely indicated. It is also possible that the formation may contain a number of intraformational overlaps that are at least partly responsible for the rapid thinning toward the east of the conglomerate occurring along Currant

and Red creeks. It seems more likely, however, that the conglomerates were deposited in alluvial fans at the mouths of eastward-flowing streams and that the limits of the fans reached very little farther east than the present position of Duchesne River. No doubt at the same time flood-plain deposits were being laid down farther east, but none has yet been identified.

The contact between the top of the Currant Creek formation and the overlying Uinta (?) formation is obscure. The base of the Uinta (?) "red beds" may be determined in a general way by color and lithologic character, but this contact may best be seen from a distance as shown in Plate 6. Although the dips at the contact seem concordant, the Uinta (?) formation undoubtedly unconformably overlies the Currant Creek.

AGE AND CORRELATION: No fossils could be found in the Currant Creek formation. Jacobsen (1940), however, found a boulder containing Upper Cretaceous fossils in one of the conglomerate beds about 600 feet above the base of the formation along the Duchesne River. One of the fossils was identified as *Placenticeras whitfieldi*, a cephalopod of Montana age.

Since the Currant Creek formation cannot be traced far from the general area of the south flank of the Uinta Mountains, no direct correlation can be made between it and any other formation in the region. By using lithologic characteristics, stratigraphic position, and other general relationships, however, a tentative correlation can be made as follows:

The formation does not transect rocks younger than Niobrara, although Niobrara and much younger rocks are found to the north in Wyoming, to the south in the Book Cliffs region, and to the east in the Asphalt Ridge vicinity. Where it overlies Niobrara rocks it does so without apparent discordance of dip, indicating that locally no great orogenic epoch preceded its deposition. The greatest mountain-building action apparently came after deposition of its basal conglomerate and probably continued while the formation was being laid down. This conclusion is suggested by the mountainward thinning of individual strata (Pl. 6) and the great divergence of dips at the top and bottom of the formation. The Currant Creek formation, therefore, must be regarded as a series of transitional strata that were being laid down from late Upper Cretaceous through Paleocene to possibly early Eocene time.

At the north end of the Wasatch Plateau, about 30 miles to the south, Spieker (personal communication) has found that the Price River formation (Montanan age) of the Mesaverde group changes facies rapidly from sandstone to conglomerate, going westward, and overlies unconformably the lower members of the Mesaverde group (Pl. 3). The North Horn formation, consisting mainly of shale and sandstone, conformably overlies the Price River formation. According to Spieker (1935, p. 106) it is Upper Cretaceous at the base and Paleocene at the top. Concordantly overlying the North Horn are the Flagstaff limestone of probable Paleocene age and the Colton formation of probable Eocene age.

The conglomerate beds of the Currant Creek formation resemble the conglomeratic facies of the Price River formation as it is exposed in the "Narrows" of Spanish Fork canyon. Also, the upper sandy beds of the Currant Creek are similar lithologically

to the North Horn formation at the same locality. It seems probable, therefore that the Currant Creek formation, because of its similar lithologic character and position above a regional unconformity in the Upper Cretaceous rocks, is roughly equivalent to the strata represented by the Price River and North Horn formations. (Pl. 3).

In southwestern Wyoming, conglomeratic strata of the Evanston and Almy formations have great thickness and unconformably overlie the coal-bearing Adaville formation of Upper Montanan age. Veatch (1907, p. 76–90) calls the Evanston upper Laramie in age and places the Almy at the base of the Eocene, although no apparent unconformity separates them. The writer suggests that further work might show a correlation between these beds and the Currant Creek formation.

CONDITIONS OF ORIGIN: The lithologic characteristics of the Currant Creek formation indicate that it was deposited at or near the mouths of streams and rivers flowing out of mountains to the west. Large alluvial fans composed of coarse pebbles and boulders came into existence at these points of debouchure. Farther east, on a river flood plain, the finer muds and sands now forming the sandy shale and sandstones of the formation were deposited. Bodies of water with sandy beaches existed on this flood plain, for ripple-marked sandstones occur at the base of the section.

TERTIARY SYSTEM

EOCENE SERIES

Wasatch formation.—The only known exposure of the Wasatch formation in the Uinta Basin is at the extreme eastern end where it forms a prominent cuesta known as Raven Ridge (Pl. 7). This outcrop was first described by White (1877, p. 277–279) of the Hayden expedition. Gale (1910, p. 75–79) mapped it later.

The formation consists of whitish-gray, pinkish, and green soft shales interstratified with thin-bedded brown sandstones. Some paper shales and marly beds are also present. The base of the formation is indefinite, as it is similar lithologically to the underlying upper Mesaverde strata, and no discordance of dip is apparent. The base was taken at a sandy shale bed containing numerous *Goniobasis*-like fresh-water gastropods.

The upper boundary of the formation is the base of the Green River shale, and it also is not sharply defined. There is a gradation between the Wasatch variegated shales and the whitish-gray shale of the Green River. On the basis of lithologic changes a thickness of 780 feet was measured. Farther west the Wasatch is completely overlapped by the Green River (Pls. 1, 2, 7). The presence of the paper shales, the fine-grained, well-sorted sandstone, and the beds of *Goniobasis*-like gastropods seems to indicate that the formation is typically of lacustrine origin and was probably deposited as mud and fine sand in a fresh-water lake.

In the extreme southwestern portion of the Uinta Basin, the Colton formation is exposed. As determined by Spieker (personal communication), it is probably of Wasatch age and is so indicated on Plate 2. It is equivalent in age to a portion of the Green River formation with which it interfingers on its upper boundary.

The Colton consists of reddish-pink sandstones and shales and forms the base of

the Roan Cliffs from Price River to the Roan Plateau in Colorado. It varies greatly in thickness. Near Soldier Summit it disappears into the Green River formation. Above the town of Green River, Utah, it is 3400 feet thick. It is not known to outcrop on the south flank of the Uinta Mountains. Further work may show a correlation between the lower part of the Uinta (?) formation and the Colton since their lithologic character and stratigraphic position are similar.

Green River formation.—Most of the southern part of the Uinta Basin consists of outcrops of the Green River formation. A thin band of it is also exposed on the south side of the Rim Rock from near the Colorado border to Asphalt Ridge (Pl. 7).

The formation has been studied in this region by Winchester (1918), Woodruff and Day (1914), Sears and Bradley (1924), Bradley (1931), and others.

It is characteristically composed of greenish-gray and white clay shale. There are occasional beds of hard blocky sandstone, but they rarely exceed a few feet in thickness. Paper-thin elastic shales are frequently seen in exposures. Oil shales and bituminous sandstones are present in some places, usually near the middle of the section.

The formation varies considerably in thickness. Along Indian Canyon, above Duchesne, it exceeds 5000 feet, while farther east at the Duchesne County line about 1800 feet is present. At the Colorado border the writer measured 2400 feet.

Along the north side of the Uinta Basin, the lower boundary of the Green River shale is a surface of overlap. From Asphalt Ridge, eastward along the Rim Rock, these beds unconformably overlie the Mesaverde group (Pls. 1, 7). Throughout this distance the base of the Green River rises stratigraphically until, at Raven Ridge, the Wasatch is exposed. The Wasatch formation is overlapped by the Green River but is completely concordant with that formation.

At the south side of the Uinta Basin, the Green River shale overlies and interfingers with the Colton formation. It also concordantly overlies the Flagstaff limestone at the north end of the Wasatch Plateau.

Throughout the south and middle portions of the Basin, the Green River strata are conformably overlain by the Uinta formation. However, everywhere that it is exposed on the north side of the Basin, the top is marked by an unconformity. The Duchesne River formation, from Vernal east to Colorado, unconformably overlies these shale beds. The reddish-orange shales of the Duchesne River strata contrast strikingly with the gray-green Green River shale, making the contact distinct.

The manner in which the 5000 feet of Green River shale exposed in Indian Canyon, above Duchesne, disappears and the reason for its absence from the outcrops on the south flank of the Uinta Mountains, only 12 miles distant, is still unknown. Possibly that part of the formation mapped as Uinta (?) along the Duchesne River, measuring 5400 feet thick and thus representing a seemingly impossible threefold increase in thickness of the Uinta formation only 2 miles distant, may be a tongue of variegated shale or change of facies extending into the main body of the Green River formation. The Uinta (?) formation lies unconformably on beds of various ages along the north side of the Basin (Pls. 1, 2, 9) and it is also possible that this relation accounts for the absence of the Green River shale there.

Uinta formation.—The central portion of the Uinta Basin, just south of the syn-

clinal axis, is occupied by outcrops of the Uinta formation. These beds were first described by Osborn (1894), and much paleontological work has been done since Riggs (1912) and Douglass (1914). Kay (1934, p. 347–372) has done the most recent and detailed work on the formation.

The Uinta formation of the central part of the Basin consists of variegated shale interbedded with gray and buff sandstone. Kay's measurements showed 1800 feet from Duchesne to Rock Creek and 1648 feet from Ouray north. The Uinta lies conformably upon the Green River shale and is overlain conformably at these localities by the Duchesne River formation.

The lithologic character and fossils of the formation indicate it to be a fluviatile deposit.

Uinta (?) formation.—A sequence of variegated shales and conglomeratic beds outcrop as a prominent hogback from Deep Creek to Whiterocks River. Kay (personal communication) called these beds Uinta on the basis of their lithologic character. Lupton (1912, p. 9) called them Wasatch. Because these beds have not been positively identified the writer refers to them as the Uinta (?) formation. At Deep Creek they unconformably overlie the Mancos shale and dip 30°S. (Pl. 10, fig. 1). The much flatter-dipping Duchesne River formation is distinctly unconformable upon these beds.

Along the Duchesne River thick "red beds" overlap the Currant Creek formation and are overlain unconformably by the Duchesne River formation. They are shown on the map (Pl. 2) as Uinta (?) formation. Their thickness is 5400 feet. These beds may be traced east to Rock Creek, where Kay (personal communication) found a titanothere jaw in them west of Mountain Home, and on this and lithologic evidence he considered them to be the Uinta formation.

If they are Uinta, it is difficult to account for the fact that in the 2 miles along the Duchesne River, where they are covered by the Duchesne River formation, they thicken from 1800 feet to 5400 feet. Also, their lithologic character changes from reddish-brown, gray, and green shales with light-brown and gray sandstone to red and white shales interbedded with buff sandstones. For this reason, the writer believes that, while the top of these beds is no doubt Uinta, the lower part may be equivalent to either the Wasatch or Colton formations of the north end of the Wasatch Plateau. It also seems possible that a part of the Uinta (?), as mentioned above, may be a facies change of the Green River formation similar to the Cathedral Bluffs tongue of northwestern Colorado.

Plate 2 and Plate 9 show how the Uinta (?) formation is bounded on top and bottom. It plainly overlaps both the Mesaverde group and the Currant Creek formation and is in turn overlain unconformably by the Duchesne River formation.

OLIGOCENE SERIES

Duchesne River formation.—This formation outcrops along the axis and the north side of the Uinta Basin. Formerly considered upper Uinta, it was studied and described by Peterson and Kay (1931, p. 293–305), but later Kay (1934, p. 357–371) referred it to the Oligocene.

Characteristically this formation is composed of reddish-orange shales, buff to

FIGURE 1. CRETACEOUS SECTION AT THE UTAH-COLORADO BORDER
Looking west to Blue Mountain. Weber formation forms the distant cliffs.

FIGURE 2. UINTA (?) FORMATION UNCONFORMABLY OVERLYING THE MANCOS SHALE
Looking west at Deep Creek.

CRETACEOUS SECTIONS AT THE UTAH-COLORADO BORDER AND AT DEEP CREEK

gray sandstones, and conglomerates in the eastern part of the Basin. In the western portion of the area along the Duchesne River, it consists mainly of massive light-gray to buff sandstones interbedded with sandy brown and red shales. Kay's (1934, p. 362–370) measurements show more than 1500 feet along his Rock Creek traverse and 1372 feet north of Ouray.

The Duchesne River formation lies conformably on the Uinta formation throughout the axial region of the Basin. However, its northern boundary shows that it unconformably overlies strata which range from Carboniferous to Eocene in age (Pl. 2). At Farm Creek Peak (Pl. 9), it overlies the Uinta (?) unconformably.

At Asphalt Ridge, there is an interesting unconformable contact between the Mesaverde group and the Duchesne River formation. The Cretaceous sandstones were truncated by erosion representing a considerable time interval, for about 7000 feet of strata is missing. The Duchesne River formation was deposited on this surface. The sandstone beds near the base of the Duchesne River, as well as the Mesaverde sandstones, are saturated with bituminous material, indicating introduction after deposition.

The location and lithologic character of the Duchesne River show that its sediments came almost entirely from the Uinta Mountains to the north.

MIOCENE (?) SERIES

Bishop conglomerate.—Considerable field work has been done on this formation by many workers, and their opinions as to its age differ somewhat. The scope of the writer's work precluded any detailed work on this formation, so he has tentatively called the beds Miocene (?). It has since been suggested by Kay (personal communication) that they are Pleistocene because he found Pleistocene fossils in the conglomerate and associated material in Colorado. This evidence seems the most conclusive offered thus far.

The Bishop conglomerate rests with angular unconformity upon strata of pre-Cambrian to Oligocene age. Because it lies at high elevations where the vegetation is profuse, the details of its lithologic character and attitude of strata are not always apparent. Also, the pebbles from the weathered conglomerate are not readily distinguishable from the numerous glacial gravels of the area.

The conglomerate probably represents debris deposited on high-level erosion surfaces or pediments.

QUATERNARY SYSTEM

Pleistocene glacial drift and its associated debris are common along the north edge of the Uinta Basin where the various river valleys emerge from the higher mountains. Neither these deposits nor the recent river gravels and valley alluvium have been shown on Plate 2.

IGNEOUS ROCKS

A small area of volcanic agglomerate and andesitic extrusives is shown on the extreme northwestern portion of Plate 2. According to Forrester (1937, p. 642) and others these igneous rocks rest on Eocene sediments and are pre-Miocene, hence must be late Eocene or early Oligocene.

STRUCTURE

GENERAL FEATURES

The Uinta Basin of Utah is a broad asymmetrical syncline. Its axis trends roughly east but bends gently toward the north. The syncline pitches steeply toward the west near Currant Creek and gently toward the southeast on Deadman Bench, north of Bonanza. The topographic low of the Basin is not coincident with the axis of the syncline but parallels it throughout its length about 10 to 15 miles to the south.

The south side of the Basin is bounded by the San Rafael swell and the Uncompahgre uplift, and low dips, seldom over 6°, predominate. South of Myton, a re-entrant into the south limb of the syncline results from a further flattening of the dips in that locality (Pl. 2).

The north side of the Uinta Basin syncline is also the south flank of the Uinta Mountain anticline, and the axes of the two folds are essentially parallel. Overlaps by Tertiary rocks prevail along this flank. The dips of the pre-Eocene rocks on this side are much steeper than those on the south so the syncline is distinctly asymmetrical.

Two prominent re-entrants into the Uinta Mountain anticline are present on the north side of the Basin. One extends from Rock Creek to Uinta River; the other lies northeast of Vernal between Split Mountain and the main Uinta range. Two steeply dipping noses, Split Mountain and Blue Mountain, supplementary and parallel to the main anticline, plunge westward into the Basin in the vicinity of Jensen. Several smaller folds with north-south axes plunge into the synclinal basin along the south flank of the Uinta anticline.

Faulting is probably not important in the structure of the Uinta Basin.

FOLDING

General considerations.—While the Uinta Basin syncline is not crossed by any prominent belts of folding, it is not in all respects a completely simple fold. Its northern flank is crenulated by numerous flexures, some belonging to the main Uinta Mountain anticline, and others representing supplementary structures on the flank of the major fold. Forrester (1937, p. 631–666) has discussed the larger features associated with the Uinta Mountain anticline. Only the structures more intimately related to the Uinta Basin syncline will be described in this report.

Vernal Valley trough.—The wide Mancos shale re-entrant northeast of Vernal represents a local downwarping between the rising masses of the main Uinta anticline and the Split Mountain-Blue Mountain plateau. It pitches steeply at the northeast end but flattens completely and then reverses its pitch north of Vernal to form a terrace-like structure. The steep westward-pitching Split Mountain and Blue Mountain noses, which are supplementary folds on the main Uinta anticline, plunge into this terrace. Their structural trends die out east of Asphalt Ridge, for their continuation cannot be recognized in the attitude of the strata there. Probably the local structural high responsible for the accumulation of gas in the Ashley Valley field is partially related to the Blue Mountain nose.

Rangely anticline.—While only a small portion of this fold lies in Utah, it is of re-

gional interest. Gale (1908) considered it a supplementary fold of the Uinta Mountain anticline because the axes of the two are essentially parallel. Forrester (1937, p. 653) concurred. While this and other similarities in structural attitude indicate they are the results of similarly oriented stresses, other field evidence points strongly to the improbability of their contemporaneity.

On the flanks of the Rangely anticline the Green River beds grade downward imperceptibly into the Wasatch, and the latter in turn rests disconformably on the Mesaverde, even where the dips are as steep as 42°. Along the Rim Rock hogback, however, the Wasatch is missing, and the Green River overlies the Mesaverde strata in angular unconformity. Apparently there were two periods of folding—one pre-Wasatch, the other post-Green River.

Neal nose.—Northwest of Vernal is a nose or plunging anticline with the axis at right angles to the main trend of the Uinta Mountain anticline. Locally this structure is called the Neal "dome", and it has been tested for oil. While the dips are low the structure is expressed in the topography by several hogbacks.

Deep Creek nose.—A nose with a north-south axis lies along Deep Creek. The basal conglomeratic strata of the Uinta (?) formation give topographic expression to the fold in the form of a well-marked curving hogback that outcrops from Little Mountain to Mosby Mountain. This hogback parallels the Frontier sandstone cuesta throughout this distance (Pl. 10, fig. 1), indicating that the structure did not come into existence until after deposition of the Uinta (?). Furthermore, the divergence of dip between the Mancos shale and the Uinta (?) at their contact is only 5°. The sections on Plate 2 show the details of this structure.

Whiterocks River nose.—A small steeply dipping nose with a north-south trend pitches into the Uinta Basin near Whiterocks River. Large quantities of bituminous material are present in the Jurassic sandstones at this locality, and the accumulation is probably due to this structure.

Transverse folding.—Re-entrants on either flank of the basin seem to indicate that a minor transverse syncline probably cuts across the basin from the vicinity of the Uinta River to the region south of Myton. This is the only field evidence that the Uinta Basin may be crossed by minor belts of folding.

FAULTING

The only fault of importance affecting the Cretaceous rocks of the basin is at the north end of Asphalt Ridge. It is a normal fault and has a throw of less than 250 feet. It is post-Oligocene, for the Duchesne River formation is displaced.

STRUCTURAL HISTORY

GENERAL STATEMENT

The Uinta Basin is not the result of a single diastrophic event but a series of events, both orogenic and depositional. While field evidence is not overabundant, the general sequence is fairly plain and is as follows:

DISTURBANCE IN THE TABBY MOUNTAIN AREA

Formation of the Uinta Basin began with major diastrophism, the uplift and erosion of Coloradoan and probably lower Montanan rocks in the region around

Tabby Mountain. This disturbance is represented by the unconformity between the Currant Creek formation and the Mesaverde group of sandstone of Niobraran age (Pl. 1). The concordance of dips on either side of the contact indicates uplift without tilting. The westward source of the Currant Creek conglomerate suggests newly uplifted mountains in that direction. This orogeny represents the initial phase of the Laramide Revolution in this area. Spieker (1936, p. 62–63) has shown that folding of the Wasatch Mountains near the north end of the Wasatch Plateau, 30 miles to the south, preceded the deposition of the conglomeratic facies of the Price River formation (upper Montanan age). Since the Currant Creek and Price River formations appear to be approximately correlative, probably these two movements were contemporaneous.

Uplift and folding evidently did not take place in the east end of the Uinta Basin at this time, or erosion would have cut into the Coloradoan rocks as it did in the west (Pl. 1). Likewise, Spieker (1935, p. 106), in his work on the Wasatch Plateau, found that this primary orogeny, although the most important in the folding of the Wasatch Mountains, did not extend eastward into the Book Cliffs.

FOLDING OF THE UINTA MOUNTAIN ANTICLINE

The first diastrophism involving the whole Uinta Basin was probably contemporaneous with deposition of the Currant Creek formation. This formation everywhere shows a 25- to 30- degree divergence of dip between its top and bottom, indicating constant movement during deposition. Plate 6 shows the resulting mountainward thinning of these beds due to continuous uplift. Under such conditions the Uinta anticline rose to a considerable extent without interruption of sedimentation. Such deposition could have been possible only where excess debris was available, such as near the base of a mountain range of considerable relief. Since a general westward source of the Currant Creek conglomeratic beds is certain (Pl. 1), and the quartzite and chert pebbles of which it is composed could have come only from Paleozoic or older rocks, the Wasatch Mountains, immediately to the west, seem definitely to have been the source. In addition, numerous cut and fill channels in the conglomeratic beds all show a general east-west alignment. Since the Uinta Mountain anticline did not contribute much sediment to the Currant Creek formation, it must have had little relief during this long period of folding until near the end of the orogenic epoch, when the basin in which the Wasatch and Green River formations were deposited came into existence. This final movement is shown by the unconformable contact between the Tertiary and Upper Cretaceous along the Rim Rock hogback. Even then, however, the amount of relief the Uinta Mountains attained was no doubt small, as compared with that at the present time. If the Currant Creek is equivalent to the Price River and North Horn formations of the Wasatch Plateau, this period of warping began in upper Montanan time and extended into the Paleocene.

Eastward along the Uinta anticline, the amount of warping must have progressively lessened because at the east end of the anticline in the Axial Basin no interruption occurred throughout the sedimentation of the Williams Fork, Lewis shale, and the Laramide group (Sears, 1924, p. 290–292). The progressive west to east

nature of the Laramide folding and the attendant rise in stratigraphic section of the major unconformity associated with this orogenic epoch has been recognized throughout the Rocky Mountain region. Bartram (1939) described it very well. It is shown typically in the Uinta Basin (Pls. 1, 3).

FIRST UPLIFT OF THE UINTA MOUNTAINS

Following deposition of the Green River shale, the Uinta Mountains must have been uplifted somewhat to produce the unconformity below the boulder conglomerate at the base of the Uinta (?) formation where it overlies the Jurassic and Cretaceous rocks in the Whiterocks River-Deep Creek area (Pl. 1).

This same uplift and the attendant great quantities of sediment being deposited probably account for the overlap of the Uinta (?) formation on the Currant Creek beds, Mesaverde group, and Mancos shale in the Duchesne River-Rock Creek vicinity (Pl. 2).

SECOND UPLIFT OF THE UINTA MOUNTAINS

Following deposition of the Uinta (?) formation, probably at the close of the Eocene, the second and greatest uplift of the Uinta anticline occurred. Forrester (1937, p. 48) described this uplift and related it to the emplacement of a batholithic core in the main Uinta anticline. This orogeny tilted the Uinta (?) formation to angles of 36° in the Whiterocks River-Deep Creek area. On the truncated surfaces of these and older upturned beds, the Duchesne River formation was deposited. Later erosion has stripped away much of the Duchesne River cover.

In the Tabby Mountain area, the thick Uinta (?) formation was tilted by this uplift, and the Duchesne River formation was deposited on the truncated edge of the beds. It overlaps the 5400 feet of Uinta (?) formation at Farm Creek Peak (Pl. 9).

From Asphalt Ridge eastward into Colorado, erosion following the uplift removed the Uinta formation, part of the Green River shale, and a part of the Mesaverde group from the north side of the basin. The Duchesne River formation was laid down on this erosion surface.

MOVEMENT OF POST-DUCHESNE RIVER AGE

The Uinta Basin acquired its present structural form after deposition of the Duchesne River, for the formation was involved in the folding. Probably, the rising at this time of the San Rafael Swell and the Uncompahgre Plateau on the south determined the axis of the basin. Movement occurred along the north side of the basin also for, in some places there, the Duchesne River strata are inclined at angles greater than initial dips for the type of sediments concerned. Moreover, from Asphalt Ridge to Little Mountain, dips in this formation are as high as 23°S.

SUBSIDENCE OF THE UINTA ANTICLINE

After deposition of the Bishop conglomerate, the Uinta range subsided. Sears (1924, p. 287) recognized this movement on the north side of the mountains. The evidence can be seen at several places on the south flank of the Uintas. The Duchesne River formation on the top of Little Mountain shows a distinct northward dip. At Farm Creek Peak a local flattening of the dip of the same formation indicates settlement of the mountain mass on the north (Pl. 9).

Subsidence has locally produced small steeply dipping flexures that obviously are not related to the main structural features of the area. Such a settlement is clearly expressed in the topography east of Whiterocks where the Duchesne River formation dips 13°N.

CONTINENTAL UPLIFT

It is generally accepted that an epeirogenic uplift took place throughout the Rocky Mountain region in late Tertiary or early Quaternary time. An uplift of from 8000 to 10,000 feet was estimated by Forrester (1937, p. 631–666) for the Uinta Mountains.

The only direct evidence of this uplift is shown by the entrenchment of the meanders of the Green River system. This feature has been described in numerous publications.

MIGRATION OF THE SYNCLINAL AXIS

The progressive southward thickening of the Wasatch and Green River formations indicates that the deepest part of the Uinta Basin, and therefore probably the synclinal axis, was once farther south. Also, the present position of the Strawberry-Duchesne River system almost exactly parallels the synclinal axis of the Basin but is displaced 5 to 10 miles to the south. Admittedly, this evidence is meager, but it suggests that the axis of the Basin may have migrated northward during Tertiary deposition and orogeny.

SUMMARY OF CONCLUSIONS

The Cretaceous rocks of the Uinta Basin are present along the south flank of the Uinta Mountain anticline and are overlain unconformably by Tertiary sedimentary formations throughout this area.

The basal Upper Cretaceous Dakota (?) sandstone lies disconformably on the Jurassic Morrison and is overlain conformably by the Mancos shale. It is suggested that the thick basal conglomerate bed in the Dakota (?) at the Colorado border is Lower Cretaceous.

The Mancos shale refers only to a lithologic unit, since in the Vernal vicinity it is equivalent stratigraphically to the Mesaverde group sandstones of the Tabby Mountain region to the west. It is divided into the following members, all easily identified throughout the Basin: lower shale member, Aspen shale, middle shale member, Frontier sandstone, and upper shale member.

The Mesaverde group is mapped only as a lithologic facies, since at the west end of the area it is predominantly Niobrara in age, whereas at the east it is completely post-Niobrara. In the Vernal region, it is divided into three members: the Asphalt Ridge sandstone, the Rim Rock sandstone, and the Williams Fork formation. The lower members of the Mesaverde group intertongue with the upper portion of the Mancos shale and wedge out toward the east.

The author proposes the name Currant Creek for a sequence of conglomeratic beds unconformably overlying the Mesaverde group in the Tabby Mountain region. This new formation is referred tentatively to the Upper Cretaceous. It probably represents the transitional strata deposited from late Upper Cretaceous to Eocene time during the early phases of the Laramide folding.

The Uinta Basin is an asymmetrical syncline with numerous small noses and supplementary folds plunging southward into it from the Uinta Mountain anticline on the north. Its structural history involves the following movements: local uplift, folding, two periods of uplift, regional tilting, subsidence, continental uplift.

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